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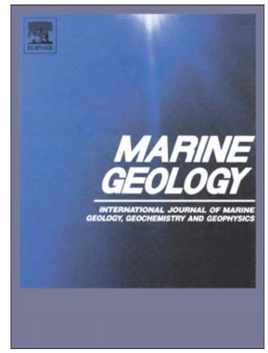

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Oceanographic processes and morphosedimentary products along the Iberian margins: A new multidisciplinary approach

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Abstract

Our understanding of bottom-currents and associated oceanographic processes (e.g., overflows, barotropic tidal currents) including intermittent processes (e.g., vertical eddies, deep sea storms, horizontal vortices, internal waves and tsunamis) is rapidly evolving. Many deep-water processes remain poorly understood due to limited direct observations, but can generate significant depositional and erosional features on both short and long term time scales. This paper represents a review work, which describes for the first time these oceanographic processes and examines their potential role in the sedimentary features along the Iberian continental margins. This review explores the implications of the studied processes, given their secondary role relative to other factors such as mass-transport and turbiditic processes, and highlights three major results: a) contourite depositional and erosional features are ubiquitous along the margins, indicating that bottom currents and associated oceanographic processes control the physiography and sedimentation; b) the position of interfaces between major water masses and their vertical and spatial variation in time specifically appears to exert primary control in determining major morphologic changes along the slope gradient, including the contourite terraces development; and c) contourites deposits exhibit greater variation than the established facies model suggests. Therefore, a consistent facies model however faces substantial challenges in terms of the wide range of oceanographic processes that can influence in their development. An integrated interpretation of these oceanographic processes requires an understanding of contourites, seafloor features, their spatial and temporal evolution, and the near-bottom flows that form them. This approach will synthesize oceanographic data, seafloor morphology, sediments and seismic images to improve our knowledge of permanent and intermittent processes around Iberia, and evaluate their conceptual and regional role in the margin's sedimentary evolution. Given their complexes, three-dimensional and temporally-variable nature, integration of these processes into sedimentary, oceanographic and climatological frameworks will require a multidisciplinary approach that includes Geology, Physical oceanography, Paleoceanography and Benthic Biology.

Key words: Iberian margins, oceanographic processes, bottom currents, sedimentary processes, contourites

1. Introduction

Deep-marine settings have been considered until 1980s as relatively low energy and quiescent depositional environments where deep-water masses flow as relatively slow-moving tabular bodies and deposition is episodically interrupted by down-slope gravity driven processes. However, since the 1990s it has been demonstrated that deep-water masses can exhibit relatively high velocity and play a dominant depositional role in certain areas (e.g., Rebesco *et al*., 2014). These conditions hold especially when water masses interact with local seafloor irregularities (i.e., seamounts, ridges, hills, mounds, banks, scarps, etc.). Flow through deep oceanic gateways, straits or basins can also generate high velocity and turbulent flows, resulting in regional-scale erosional, depositional and mixed morphological features. Understanding these seafloor features therefore requires a more detailed model of bottom current effects.

The general term 'bottom current' refers to deep-water currents capable of eroding, transporting, and depositing sediments along the seafloor (Rebesco and Camerlenghi, 2008). Background bottom currents are the result of both the ocean thermohaline circulation (THC) and the ocean wind-driven circulation (Rahmstorf, 2006). They generally exhibit persistent net flow along-slope (following the local bathymetry), but can considerably vary in direction and velocity (Stow *et al*., 2009). Overflows, tides, including internal tides and other intermittent processes such as bottom reaching (vertical) eddies, deep sea storms, (horizontal) vortices, internal waves, solitons, tsunami related currents, rogue waves and cyclonic waves (during storms and hurricanes) can also affect bottom currents (Fig. 1) (e.g., Shanmugam, 2012a, 2013a, b, 2014; Rebesco *et al*., 2014). These processes modulate the bottom currents and their speed, instantaneous direction, and tend to develop local and regional hydrodynamic structures (e.g., cores, branches, filaments, eddies, vortices, local turbulence, internal waves, helicoidal flows, vertical columns, etc.). Many of these oceanographic deep-water processes are poorly understood but can generate pervasive depositional and erosional features over both short- and long- term time scales (Shanmugam, 2013b; Rebesco *et al*., 2014 and reference therein).

Contourites are defined as sediments deposited or substantially reworked by the persistent action of bottom currents (e.g., Stow *et al*., 2002a; Rebesco, 2005, Rebesco *et al.,* 2014). The term "contourites" was originally intended to define sediments deposited in the deep sea by contour-parallel thermohaline currents (Hollister and Heezen, 1967). Usage of the term has subsequently widened to include a larger range of deposits affected by different types of currents (Rebesco *et al*., 2014). Bottom currents are capable of building thick and extensive accumulations of sediments referred to as "contourite drifts". The term "contourite depositional system" (CDS) refers to large contourite deposits (drifts) associated with erosional features in an environment dominated by along-slope processes due to a particular water mass (Hernández-Molina *et al*., 2003, 2008a,b, 2009).

The Iberian continental margins have had a very complex and varied origin as well as geodynamic and sedimentary evolution. For decades, many studies have been carried out on the structure and evolution of these margins (e.g., Maestro *et al.*, 2015 and reference therein). The circulation of water masses around Iberia leads to the development of bottom currents and rather unknown associated oceanographic processes, some of which impinge upon the seafloor with relatively high velocity and interact along the continental slope. The resulting contourite erosional and depositional features comprise extensive, complex and often poorly known CDSs in various geological settings, consituting valuable sedimentary records of margins evolution (see compilation from Hernández-Molina *et al.*, 2011; Llave *et al.*, 2015a,b). Therefore, the present work have three main objectives: (1) to combine previous work and new data on describing persistent or intermittent oceanographic processes associated with bottom currents circulation along the Iberia continental margins; (2) to evaluate their role in determining the shape, potential sedimentary products and evolution of the margins; and (3) to propose some future considerations contemplating the implications of these findings.

2. Methodology and data set

Present work constitutes mostly a review of described persistent or intermittent oceanographic processes associated with bottom current circulation along the Iberia continental margins. Such a compilation is difficult, since there is a high degree of heterogeneity in both data coverage and terminology. It represents a collaborative work achieved during the last years by Geologists, Physical Oceanographers, Paleoceanographers and Benthic Biologists. A varied data set were compiled from numerous bibliographic sources or unpublished, including: radar images; conductivity, temperature and depth (CTD) stations; Acoustic Doppler Current Profiler (ADCP); time series observations of the near-bottom current speed, potential temperature and suspended sediment concentration (SSC); submarine bottom photographs; superficial samples; swath bathymetry data; side scan sonar images; high- and low-resolution seismic reflection profiles; and sediments samples taken by drilling and numerical modelling.

A first part of the work offers a conceptual description of persistent or intermittent oceanographic processes associated with bottom current circulation inspired in recent works from Shanmugam (2012a, 2013a, b, 2014) and Rebesco *et al*. (2014), but including, by first time, an exhaustive compilation of them around the Iberian margins. The second part describes how to proceed with the unique research challenge of integrating these complex, three-dimensional and temporally variable oceanographic processes into sedimentary and climatological models.

3. Oceanographic processes

3.1. Bottom currents and thermohaline circulation

The large-scale deep-water circulation is a critical part of the global conveyor belt that regulates the Earth's climate (Rahmstorf, 2006; Kuhlbrodt *et al*., 2007). In regions where the deep ocean currents are close enough to the ocean bottom they may shape the deep seafloor. As a first order, dense water formation at high latitudes drives the global THC. Salinity or temperature driven density variations cause sinking of the surface water masses towards greater depths and control their subsequent transport and eventual global distribution. Deep-water formation is closely associated with convection in a few localities. These mainly occur in the subpolar convergence zone of the Northern Hemisphere and the convergence zone of the Antarctic polar front in the Southern Hemisphere. The principal deep-water masses forming part of the THC include the North Atlantic Deep Water (NADW) and the Antarctic Bottom Water (AABW), which spread across the deepest marine domains, as Deep Western Boundary Currents (DWBC) (Rahmstorf, 2006). Bottom circulation and distribution along the oceanic basins is conditioned by their proximity to high-latitude input sources, seafloor morphology, inter-oceanic connection via deep gateways and the Coriolis Effect (Kennett, 1982; Faugères *et al*., 1993; Rahmstorf, 2006). Other bottom currents, such as the extension of the wind-generated currents, are not linked to THC. These latter currents include the Gulf of Mexico current system (e.g., Loop Current; Wunsch, 2002) and the Antarctic Circumpolar Current (Orsi *et al*., 1995).

Bottom-water masses move generaly relatively slowly as laminar flows over their global course (< 3 cm/s). Regional and local bottom current velocities exhibit local heterogeneity due to varying seafloor stress and other current instabilities, such as mesoscale variability. Western stretches of ocean basins and topographic obstacles along the seafloor facilitate higher velocities (> 0.6-1 m/s) (Stow *et al*., 2009; Hernández-Molina *et al*., 2011). Bottom water behaviour and velocity partially affect the seafloor through lateral transport of suspended particulate matter in the water column. This material is up to 10 times higher in concentration in the deep nepheloid layer (at 50 – 200 m above the seafloor) of the Deep Western Boundary Currents compared with other oceanic areas (Rebesco *et al*., 2014). A sufficiently active bottom current acting for a prolonged period of time will thus profoundly affect the seafloor by winnowing of fine-grained sediments, yelding large-scale erosion, depositional and mixed features (e.g., Stow *et al*., 2002b; Rebesco and Camerlenghi, 2008; Rebesco *et al*., 2014). Bottom currents represent long-term hydrologic conditions on geological time scales. The establishment of the main CDSs around the world in fact coincides with the Eocene / Oligocene boundary (~32 Ma), and was more recently reactivated by THC during the Middle Miocene (Hernández-Molina *et al*., 2008a).

The topic of bottom current processes around Iberia is addressed by Hernández-Molina et al. (2011) and Llave *et al*. 2015a, b (this special issue). Along the Iberian margins, several important bottom currents shape the continental margins and abyssal plains (e.g., Iorga and Lozier, 1999; Millot, 1999, 2009, 2014; Serra *et al*., 2010a). These currents derive primarily from the Western Mediterranean Deep Water (WMDW) and the Levantine Intermediate Water (LIW) in the Alboran Sea, as well as the Mediterranean Overflow Water (MOW) and the Lower Deep Water (LDW) in the Atlantic (see the water masses compliation in Hernández-Molina *et al.*, 2011 and Llave *et al.*, 2015a, b, this issue). Well-developed CDSs around Iberia (e.g., Portuguese margin) have not been systematically interpreted according to bottom currents models and thus are not well understood.

3.2. Overflows

Overflows constitute permanent oceanographic processes related to: 1) flow over a topographic barrier from a regional basin into the open ocean (Fig. 1) and 2) open-ocean flow into an isolated regional basin. The Denmark Strait Overflow Water (DSOW); Iceland-Scotland Overflow Water (ISOW) and MOW (Price and Baringer, 1994; Legg *et al*., 2009) offer good examples of the first type of overflow. The second type of overflow occurs in association with deep gateways such as the Bruce Passage between the Weddell Sea and Scotia Sea (Legg *et al*., 2009; Lobo *et al*., 2011; García et al., 2015a, this special issue) or the overflow of AABW from the Brazil Basin into the North Atlantic Basin across the Atlantic ridge or in the Gulf of Mexico (Legg *et al*., 2009). In the aforementioned cases, a dense gravity current, carrying a particular water mass, descends the regional slope to a greater depth until it reaches density equilibrium. Entrainment of surrounding water, bottom friction and inertial accelerations modify the bottom current along the slope (Fig. S1, in *Supplementary material*). Bottom friction and the Earth rotation induce the Ekman bottom boundary layer (e.g., Pedlosky, 1996; Wåhlin and Walin, 2001), which produces a net transport to the left in the Northern Hemisphere (Fig. S1). Although the frictional transport is confined to a thin layer near the bottom (the Ekman boundary layer), it affects the whole water column asymptotically. The dense water adjusts to the divergence of the frictional transport, which acts as a horizontal diffusive process, minimizing the curvature of the dense interface. The lower (seaward) edge moves downhill as the Ekman layer expels the Ekman transport from its interior (e.g., Wåhlin and Walin, 2001). At the upper (landward) edge, the dense interface then becomes almost horizontal, with diminished geostrophic velocity and frictional transport (e.g., Wåhlin and Walin, 2001). The combined frictional effect over the entire outflow causes it to gradually widen, keeping the upper horizontal boundary at a nearly constant depth (e.g., Borenäs and Wåhlin, 2000). If these processes occur over a laterally varying bottom slope, they act to split the flow into two or more cores and branches, as suggested for example, in an interpretation of the Mediterranean Outflow (Borenäs *et al*., 2002; Serra *et al*., 2010a, b). As dense water passes from its formation site into the open ocean as an overflow, it undergoes mixing with the overlying water column. This process determines the eventual properties and volume transport of the resultant bottom currents, and triggers eddy formation (Legg *et al*., 2009).

The main overflow example in the Iberian margins stems from MOW dynamics during the late Miocene opening of the Gibraltar gateway (Duggen *et al*., 2003; Bache *et al*., 2012; Roveri *et al*., 2014). The Strait of Gibraltar, which consists of a 60 km long sill and narrow gateway (Armi and Farmer, 1988), plays a key role in water exchange between the Mediterranean Sea and the Atlantic Ocean. The excess evaporation over precipitation (0.5 to 0.8 m/yr) dramatically enhances salinity and density of the Mediterranean, relative to waters of the adjacent Atlantic Ocean (Millot *et al*., 2006). Mediterranean salinity tongues develop in the Atlantic Ocean from overflow (0.68 Sv) of dense, warm (13°C) and highly saline (37) water (Bryden *et al*., 1994, Candela, 2001; Serra *et al*., 2010a). Mediterranean Levantine Intermediate Water (LIW) and Western Mediterranean Deep Water (WMDW) mix, when they flow out through the Strait and originate the MOW that is around 2 times saltier than any other Atlantic water mass. The overflow spills over the Strait of Gibraltar sills and cascades from 300 m depth at the edge of the strait down the continental slope of the eastern Gulf of Cádiz. As it does so, it entrains the overlying, less dense North Atlantic Central Water (NACW) (Johnson and Stevens, 2000). Overflows typically occur at velocities of 50 to 100 cm/s (Legg *et al*., 2009), but in the case of MOW, the current velocity decreases from around 200 cm/s in the neighbourhood of the Camarinal sill and its western approaches (Sánchez Román *et al*, 2009; Gasser *et al*, 2011), down to 60 - 100 cm/s further to the northwest. From the first 100 km of the overflow path, volume transport increases by a factor of three to four (Serra *et al*., 2010a; Rogerson *et al*., 2012). As it descends, the overflow veers north and follows the Iberian continental margins, approaching its neutral buoyancy as it nears 8° W at about 1000 m depth (Ochoa and Bray, 1991; Baringer and Price, 1997, 1999; Käse and Zenk, 1996; Ambar *et al*., 2002, 2008; Bower *et al*., 2002; Serra *et al*., 2005, 2010a; García-Lafuente *et al*., 2009).

In the Gulf of Cádiz, friction in the bottom Ekman boundary layer and entrainment in its upper part affect the MOW plume (Baringer and Price, 1997; Gasser *et al*., 2011). The vertical distribution of the along-stream velocity comprises two zones separated by a velocity maximum (called the plume nose). They consist of an upper (interfacial) layer characterised by sharp velocity and salinity gradients, and a bottom layer characterized by constant salinity and a sharp decrease in velocity (Johnson *et al*., 1994a, b). Secondary across-stream circulation spreads water from the upper layer upslope and dense water from the bottom layer downslope. In the eastern Gulf of Cádiz, near the Strait of Gibraltar, this overflow is associated with sandy-sheeted drifts, scours, ripples, sand ribbons and sediment waves (Nelson *et al*., 1993), and generates large terraces and erosive channels (Fig. 2) (Hernández-Molina *et al*., 2014). The secondary circulation here is especially relevant to the erosion, resuspension and deposition of sediment, as well as to the formation of large contourite features (Hernández-Molina *et al*., 2014).

Other examples of open-ocean overflow could happen between oceanic basins with different depths (Legg *et al*., 2009). A possible case to be explored along the Iberian margins could theoretically develop through the Theta Gap, a passage formed between La Coruña and Finisterre structural highs (Fig. 3), and narrows to approximately 5 km wide between the Biscay abyssal plain (average water depth of 5100 m) and the Iberian abyssal plain (about 5300 m deep), or other deep channels between the abyssal plains. Heezen *et al*. (1959) first described the Theta Gap connection as “a constricted passage connecting two abyssal plains which, in the vicinity of the gap, lie at different levels”. The Biscay abyssal plain regionally slopes in a southward direction, with its deepest section in the southwest corner, where the slope connects to the Theta gap. Active erosion occurs about 15 km upslope of the narrowing between the structural highs (Laughton, 1960, 1968), and creates initial incisions of up to 300 m and later incisions of about 100 m at points further along the current’s course. The LDW water mass, which makes up this overflow (>3450 m water depth), penetrates into the North Atlantic through the Discovery Gap on the western flank of the Madeira–Tore Rise (Haynes and Barton, 1990; McCartney, 1992; Van Aken, 2000), and veers northward along the Galicia margin as a near-bottom flow (Paillet and Mercier, 1997; Van Aken, 2000). A cyclonic recirculation cell occurring above the Biscay Abyssal plain deflects the current poleward however, with a velocity near the continental margin of 1.2 (± 1.0) cm/s (Dickson *et al*., 1985; Paillet and Mercier, 1997) (Fig. 3).

3.3. Processes at the interface between water masses

A pycnocline represents a layer with maximum gradient of density, which can be sharp and well defined, or diffuse with a gradual transition from one water mass to the other (Fig. 1). Turbulent mixing of water masses caused by tides (or other processes) can disrupt the pycnocline, whereas stratifying processes (e.g., regional positive buoyancy flux at the surface) maintain it. The relative balance between these two factors defines the structure of the pycnocline in different regions and at different times. The interface often tilts in one dominant direction (e.g., Reid *et al*., 1977) but can be locally and temporarily displaced by eddies (e.g., Piola and Matano, 2001; Arhan *et al*., 2002, 2003) and internal waves. Energetic current patterns associated with internal waves and eddies at the interface between water masses (Reid *et al*., 1977) strongly affect the seafloor (e.g., Hernández-

Molina *et al*., 2009, 2011; Preu *et al*., 2013) through erosion and re-suspension (Dickson and McCave, 1986; Cacchione *et al*., 2002; Puig *et al*., 2004; Shanmugam, 2013a, 2014).

Some water masses circulating around the Iberian margins have remarkable density contrasts (Fig. 4 and 5). The exchange flow through the Strait of Gibraltar exhibits these contrasts for example, as do the base and top boundaries of the MOW (Serra *et al*., 2010a), which enhanced during cold (i.e.; glacial) periods (Schönfeld *et al.,* 2003; Voelker et al., 2006; Toucanne et al., 2007; Rogerson et al., 2012). The associated currents indicate a relationship between the position of the interface and the development of large contourite terraces (Hernández-Molina *et al*., 2009, 2011; Preu *et al*., 2013; Rebesco *et al*., 2014). Several examples of this relationship have been described along the Iberian margins in the Alboran Sea (Ercilla *et al*., 2002, 2015), Gulf of Cádiz (Hernández-Molina *et al*., 2014); Portuguese Margin (Fig. 4, Pinheiro *et al*., 2010; Llave *et al*., 2015b); Galician margin (Hanebuth *et al.*, 2015); and Cantabrian Sea (Fig. 4, Maestro *et al*., 2013; Sánchez-González, 2013).

3.4. Deep-water tidal currents

Surface tides are mostly driven by the ocean's response to the gravitational fields of the Sun and the Moon. Surface tides are considered barotropic tides because their associated currents are mainly depth-independent (Fig. 1). Tidal signals consist of the superposition of several harmonics (tidal constituents) with the principle lunar semi-diurnal frequency (M_2 tidal constituent; 12.42 h) exerting the strongest global effects in agreement with the tide-generating potential theory (Lakshmi *et al*., 2000). Tidal energy may be transferred to baroclinic modes in stratified fluids, thus generating baroclinic tides (Fig. 1). Baroclinic tides can dissipate tidal energy by transferring motion along distant features. Both barotropic and baroclinic tides influence the bottom water circulation in deep-water environments (Dykstra, 2012). Tidal currents change direction periodically with time, describing an elliptical hodograph usually aligned along bathymetric contours in the ocean interior. Tidal energy tends to dissipate, and hence, to exert a stronger effect on continental slopes within submarine canyons and adjacent areas (e.g., Shepard, 1976; Shepard *et al*., 1979; Viana *et al*., 1998; Kunze *et al*., 2002; Garrett, 2003; Shanmugam, 2012a, 2013b; Gómez-Ballesteros *et al*., 2014; Gong *et al*., 2013, 2015) and within certain contourite channels (Stow *et al*., 2013a). Shanmugam (2012a) has proposed that barotropic tidal currents affect land- or shelf-incising canyons connected to estuaries or rivers. Baroclinic tide currents affect slope-incising canyons but do not clearly connect to major rivers or estuary systems. Inversion of bottom current directions due to tidal influence occurs outside these canyons (Kennett, 1982; Stow *et al*., 2013a). Deep-marine tidal bottom current velocities range from 25-50 cm/s but can reach 70-75 cm/s, and exhibit periods of up to 24 hours (Shanmugam, 2012a).

According to both *in situ* measurements (tidal gauges) and numerical models, the most important harmonic affecting the Atlantic Iberian margins is M_2 (sea-level amplitudes between 100-180 cm, Fig. 6), followed by S_2 (12.00 h, 30-50 cm) (Alvarez-Fanjul *et al.*, 1997). A typical spring/neap tidal modulation of 14.8 days (the fortnightly tidal cycle) thus affects Iberian coastal areas. The regional surface tide dynamics around the Atlantic Iberian Peninsula is part of the amphidromic system that characterises the semidiurnal tidal waves in the North Atlantic (Cartwright *et al*, 1980), where a barotropic Kelvin wave propagates northward along the western Iberian coast with a phase speed close to 900 km/h. The wave diffracts eastward in the Cantabrian Sea and its amplitude increases-g steadily shoreward. Kelvin wave dynamics predict counter clockwise tidal ellipses aligned parallel to the coast. This simple model adequately approximates observed propagation directions (e.g., the largest phase lags in the Cantabrian Sea, Alvarez-Fanjul *et al*., 1997), observed and simulated wave amplitudes, which maximize along the northern coast, and open ocean dynamics. The model, however, does not adequately describe tidal current variation in the shelf/slope region where baroclinic tides start playing a significant role and the cross-shore movement becomes increasingly important. Marta-Almeida and Dubert (2006) concluded that bathymetric irregularities exert a major influence on the velocity field wherein steepened areas of the slope amplify tidal currents, orthogonally polarising tidal ellipses along bathymetric contours. Bottom features can therefore theoretically change the S–N orientation of tidal currents. Quaresma and Pichon (2013) have recently shown that the cross-shelf velocity component over the slope and shelf can supersede the along-shelf velocity component, giving rise to clockwise rotation of tidal currents. The ellipse eccentricity decreases as a function of the slope gradient. Along the Mediterranean area of the Iberian margins, tidal currents only exert significant affects in the Alboran Sea (García-Lafuente and Cano Lucaya, 1994), where phases increase eastward. The velocities of semidiurnal components decrease from a few 10^{-2} m/s to 10^{-3} m/s in the Algero-Provençal Basin (Alberola *et al*., 1995).

Both M_2 and S_2 amplitudes increase from the Gulf of Cádiz to the Bay of Biscay, and exert the dominant influence on tidal wave amplitude and current velocity throughout the shelf (Battisti and Clarke, 1982). Across-shelf, the tidal amplitude decreases offshore with an average gradient of ~0.027 cm/km. The maximum speeds of ~ 50 cm/s and 20 cm/s (in a clockwise direction) for the M_2 and S_2 ellipses (respectively) occur on the French shelf. The Bay of Biscay confines the tidal wave and forces the generation of noticeable baroclinic tides that amplify the semidiurnal constituents as the barotropic tidal wave impinges the French Armorican shelf. Velocities are much smaller around Iberia, with maxima of 5 and 2 cm/s for the respective M_2 and S_2 ellipses (Quaresma and Pichon, 2013), and exhibit strong along-shore spatial variability. The O_1 and K_1 diurnal constituents are also present (8 and 6 cm amplitude, respectively), and cause the observed diurnal inequality of the tide.

Although the observed tidal currents are small (rarely > 20 cm/s) around Iberia, they are the primary driving force behind strong, high-frequency baroclinic activity, which does not affect other coastal regions. The combination of regional, barotropic tidal velocities and steep bathymetry amplifies this effect into "hotspots", which occur along the Atlantic margins of Iberia (Quaresma and Pichon, 2013, Fig. 6C). Hotspot areas include the Extremadura and Ortegal Spurs as well as the Gettysburg and Ormonde seamounts. The Cádiz Contourite Channel, in the central sector of the Gulf of Cádiz, offers one of the best examples of sandy bedforms created directly by the dynamic interaction of tidal currents and seafloor (Stow *et al*., 2013a, Fig. 7). The narrowness of this channel (2-12 km) amplifies bottom tidal currents due to a funnelling effect. Current meter data from this area show a strongly directional, westerly MOW moving at speeds of 1 m/s at the instrument height of 78 m with a clear periodicity in its velocity, suggestive of a semi-diurnal tidal signal and of a spring/neap cycle (Fig. 8A).

3.5. Deep-sea storms

Benthic storms (also known as deep-sea storms or abyssal storms) refer to intermittent, strong, bottom-intensified currents affecting the seafloor (Fig. 1). The HEBBLE (High Energy Benthic Boundary Layer Experiment; Hollister *et al*., 1980) first documented the existence of benthic storms while investigating the response of the cohesive seafloor to high-stress events. During these storms, bottom current velocity can increase by a factor of 2-5 times (average 15-20 cm/s with peaks of > 43 cm/s) over a period of a few days (2 to 25 days) to several weeks. Benthic storms occur at a frequency of about 8 to 10 events per year (Holger, 1987). Benthic storm related flows can rework up to several millimetres of seafloor sediment per event. Entrainment of the bottom nepheloid layer or bottom boundary layer by turbulent and diffusive mixing favours rapid resuspension of particulate matter (Gardner and Sullivan, 1981; Hollister and McCave, 1984; Nowell and Hollister, 1985; Gross and Williams, 1991). During the occurrence of these storms, the benthic boundary Layer (BBL) increases in thickness from a few meters up to 70 m, and in turbidity up to values of 5 g/l (Tucholke, 2002). Other mechanisms for benthic storms hypothesize sediment plumes descending from continental slopes and rapid input of organic-rich phytodetritus from plankton blooms in surface waters (Richardson *et al*., 1993; Puig *et al*., 2013). In this latter case, organic-rich detritus draping the seafloor is easier to flocculate and maintain in suspension. Detailed studies of suspended particulate matter throughout the BBL indicate that sediment concentration depends on the interplay of several factors, including initial concentration in suspension, turbulence, height of entrainment, velocity field, transport distance and settling velocity of the particles. Gross and Williams (1991) emphasized three physical processes, which modulate deep-sea storm events. These include turbulence in the boundary layer, eroded sediment at the sediment-water interface and suspended sediment winnowed by the flow. Any one of these three lines of evidence defines a storm event (Gross and Williams, 1991).

Benthic storms play an important role in the erosion, transport and redistribution of bottom sediments (Hollister *et al*., 1980; Nowell and Hollister, 1985; Hollister, 1993). Once disturbed by the erosional effects of increased bottom shear, sediments may be transported by bottom currents and deposited in quiet regions downstream (Gardner and Sullivan, 1981; Kennett, 1982; Hollister and McCave, 1984; Flood and Shor, 1988; Bearmon, 1989; Von Lom-Keil *et al*., 2002). These currents may travel over long distances, constrained by the Coriolis force and semi-diurnal tides parameters. Seafloor morphology from large, kilometer-scale features (e.g., seamounts) to smaller, meter-scale features (e.g., sediment waves) can affect this transport (Rebesco and Camerlenghi, 2008).

The occurrence of benthic storms relates in part to the formation of large eddies within the water column due to the interaction between major wind-driven surface currents and the bottom component of THC. Areas of the ocean with a highly variable sea surface show more frequent benthic storm activity (Richardson *et al*., 1993; Faugères and Mulder, 2011). The kinetic energy of these eddies can extend to depths of > 3000 m and thus change the magnitude and direction of bottom currents. Benthic storms also occur in the presence of strong, permanent bottom currents, and in areas where seafloor sediments are easily flocculated into suspension (Richardson *et al*., 1993). Due to the limited information concerning benthic storms, however, uncertainties persist in regards to their initiation and driving mechanism.

As major sediment entrainment processes, benthic storms play an important role in contourite development. Benthic storms also disturb benthic communities (Gross and Williams, 1991). Regions subjected to intense deep-sea storms show significant erosion of the continental slope and evidence of large submarine slides (Pickering *et al*., 1989; Gao *et al*., 1998; Einsele, 2000).

Only a few examples of deep-sea storms have been documented around Iberia linked to the formation of surface dense waters in the Gulf of Lions and the subsequent deep cascading and open-sea convection process (Font *et al*., 2007; Palanques *et al*., 2009; Durrieu de Madron *et al*., 2013). Some of these events, as the one observed in winter 2009 (Fig. 9), can be exclusively due to deep convection without the accompanying cascading process (Salat *et al*., 2010; Puig *et al*., 2012; Stabholz *et al*., 2013). In the Cantabrian margin, three independent events have been identified at La Gaviera Canyon (Sanchez *et al.*, 2014). The scarcity of information on this type of phenomena could mask their frequency. For example, a current meter moored at around 554 m water depth in the Gulf of Cádiz may have recorded possible benthic storms (Fig. 8B), but further analysis of these time series observation is required to confirm such interpretation.

3.6. Eddies

Instabilities in bottom currents often generate vortices (Serra *et al*., 2010a) and contribute to long-distance sediment transport and the formation of nepheloid layers (Fig. 1). Eddies occur when a water mass interleaves into a stratified environment, or when current flows meet a seafloor irregularity such as a canyon, seamount or cape (Roden, 1987; Rogers, 1994; Arhan *et al*., 2002; Serra *et al*., 2010a). Along some regions of the seafloor, eddy activity spans hundreds of kilometres, as for example, in the Argentine Basin (Cheney *et al.* 1983; Flood and Shor, 1988; Arhan *et al*., 2002; Hernández-Molina *et al*., 2009), the Weddell and Scotia Seas (Hernández-Molina *et al*., 2008a), the Mozambique slope (Preu *et al*., 2011), the Gulf of Cádiz and the margins west off Portugal (Serra *et al*., 2010a).

The best-studied example of eddies around Iberia relate to the MOW (sometimes referred to as meddies). Meddies are low-potential vorticity bodies that rotate clockwise and maintain MOW-type temperature and salinity conditions in their cores for long periods of time (on the order of years). These coherent vortices translate at average speeds of 2-4 cm/s and have typical azimuthal velocities of 30 cm/s (Fig. 10B). The radial structure of their core velocity resembles solid body rotation (Schultz-Tokos and Rossby, 1991). Meddies have been observed in the Canary Basin (e.g., Armi and Zenk, 1984; Richardson *et al*., 1989; Pingree and Le Cann, 1993), in the Iberian Basin (Käse *et al*., 1989; Zenk and Armi, 1990; Schultz-Tokos *et al*., 1994), off the southern coast of Portugal (Prater and Sanford, 1994; Bower *et al*., 1997; Serra and Ambar, 2002), off west Portugal (Pinheiro *et al*., 2010, Fig. 5), and along the northwestern Iberian coast (Paillet *et al*., 2002). Bower *et al*. (1997) identified Cape St. Vincent Spur and the Extremadura Promontory as points of origin for meddies. The region adjacent to Cape Finisterre (Paillet *et al*., 2002), the region of the Portimao Canyon (Serra and Ambar, 2002) and the Gorringe Bank (Serra and Ambar, 2002) are also recognized sites of meddy formation.

The proposed mechanisms for generating these eddies include boundary layer separation of the MOW undercurrent (Bower *et al*., 1997), baroclinic and/or barotropic instability of the MOW undercurrent (Käse *et al*., 1989; Cherubin *et al*., 2000), flow intermittency (Nof, 1991) and topographic effects due to capes (Pichevin and Nof, 1996; Sadoux *et al*., 2000; Serra *et al*., 2005) or canyons (Cherubin at al., 2000; Serra *et al*., 2005). Three factors contribute to the meddy displacements. These include advection by mean flow (Hogg and Stommel, 1990; Dewar and Meng, 1995), the beta-effect or meridional variation of the planetary vorticity (Nof, 1982) and interactions between different eddies (Käse and Zenk, 1996; Serra *et al*., 2002). This latter mechanism can involve the coupling of cyclones and meddies (Richardson *et al*., 2000; Carton *et al*., 2002; Serra *et al*., 2002). Near their generation sites, meddies may re-suspend sediment and carry it into the open ocean.

Eddies interaction with seafloor could produce local erosion (Hernández-Molina et al., 2008a; Rebesco et al., 2014). In this way some sub-circular depression structures and other erosive features around the Le Danois Bank (Cantabrian Sea, van Rooij et al., 2010) and the Guadalquivir Bank Margin uplift (Gulf of Cadiz, García *et al.,* 2015b, this special issue) have been attributed to the influence of local eddies of the MOW interacting with seafloor irregularities, as slide scars or other local irregularities.

3.7. Secondary circulation

Dense water masses usually have main current cores that run parallel to isobaths (Fig. 1 and 10A). The velocity of these cores manifests in contourite erosional features as moats and channels (e.g., McCave and Tucholke, 1986; Faugères *et al*., 1993, 1999; Rebesco and Stow, 2001; Stow *et al*., 2002a, 2009; Rebesco and Camerlenghi, 2008; Faugères and Mulder, 2011). They are often associated with deposition in downslope areas, and erosion in upslope areas. The structure of these features may result from helicoidal flow paths, or clockwise secondary circulation of the bottom current around the core of the current. These structures are referred to as 'horizontal eddies' in geophysical literature (Davies and Laughton, 1972; Roberts *et al*., 1974; Roden, 1987; Rogers, 1994; McCave and Carter, 1997; Hernández-Molina *et al*., 2008a, b; Zenk, 2008). Although seafloor features, as moats and contourite channels, along the Iberian margins (Garcia *et al.*, 2009; Llave *et al.*, 2015a, b) suggest helicoidal flow, this interpretation requires further analysis of currents and seafloor morphology (Rebesco *et al*., 2014). These features likely result from the Coriolis effect directing the vortex against the adjacent slope, eroding the right side of the channel and depositing sediment on the left side where the current velocity is lower (Faugères *et al*., 1999; Llave *et al*., 2007). The combined effect of bottom friction and the Coriolis effect in the Ekman layer usually results in clockwise secondary circulation in bottom currents (Wåhlin and Walin, 2001; Wåhlin, 2004; Muench *et al*., 2009; Cossu *et al*., 2010; Cossu and Wells, 2013).

This type of secondary circulation may occur in the eastern Gulf of Cádiz, near the exit of the Strait of Gibraltar (Fig. 2, Serra, 2004; Garcia *et al*., 2009; Hernández-Molina *et al*., 2014). The identification of secondary circulation further suggests its role in generating furrows observed to run oblique relative to the main bodies of the prevailing bottom-currents (Hernández-Molina *et al*., 2014).

3.8. Dense shelf water cascades

Dense shelf water cascades (DSWC; Fig. 1) describe the flow of dense water generated in shelf areas down the continental slope (Simpson, 1982; Killworth, 1983; Ivanov *et al*., 2004). DSWC are near-bottom gravity flows that develop when cooling, evaporation, freezing and/or salinization in the surface layer formed over the continental shelf causes density-driven flow over the shelf edge. The dense water plume cascades both along and across the slope area according to density, gravity effects, the Coriolis Effect, friction and mixing. Once the density of the water plume matches the density of the surrounding waters the DSWC reaches gravitational equilibrium and becomes an intermediate-depth, neutrally buoyant intrusion (Shapiro and Hill, 1997; Shapiro *et al*., 2003). Submarine canyons incised into the shelf edge can offer preferential pathways for DSWC (Canals *et al*., 2006, 2009; Trincardi *et al.*, 2007; Allen and Durrieu de Madron, 2009). This intermittent process affects biogeochemical cycles by re-suspending sediment and transporting significant volumes of minerals and organic matter (Fohrmann *et al*., 1998; Hill *et al*., 1998; Puig *et al*., 2013).

The Gulf of Lions, located in the northeastern sector of the Iberian Peninsula experiences annual DSWC (Millot, 1990, 1999; Durrieu de Madron *et al*., 2005; Puig *et al*., 2013). Winter cooling and evaporation, induced by persistent, cold, dry northerly winds (Tramontane and Mistral winds) cause increased density and mixing of coastal waters. This occurs in spite of buoyancy effects from freshwater input of the Rhone River. The prevailing westerly coastal circulation, the narrowing shelf and the topographic constraint of the Cap de Creus peninsula, all facilitate DSWC and sediment transport. These currents develop in the Gulf's southwestern sector, primarily moving through the Lacaze-Duthiers and Cap de Creus submarine canyons, where cascading flows can reach velocities > 80 cm/s (Palanques *et al*., 2006). DSWC events occur from January to April/May with duration of several days, and often begin and/or are enhanced during storms, which cause increases the suspended sediment concentration and generally amplify down-canyon sediment fluxes (Palanques *et al*., 2008). In very dry, windy and cold periods, such as the 1998-99, 2004-05 and 2005-06 winters, the Gulf developed unusually intensive DSWC that lasted for several weeks (Fig. 11). During these major events, large quantities of water and suspended particles are rapidly advected over depths of hundreds of meters along submarine canyons, causing erosion of the seafloor and generating sedimentary furrows (Palanques *et al*., 2006, 2009, 2012; Canals *et al*., 2006; Puig *et al.*, 2008). The most persistent and penetrative cascading pulses generally follow a multi-step sediment transport mechanism. Sediment is initially transported from the shelf to the upper canyon region. The subsequent cascading pulses re-suspend and redistribute the sediment down-slope, generating massive sand beds at the canyon head region (Gaudin *et al*., 2006). At depths of 1500 m, where the Cap de Creus submarine canyon widens, most of the dense water and suspended sediment leaves the canyon and flows along the northern Catalan continental margin as a contour current (Palanques *et al*., 2012). Continuous monitoring of cascading events in the Cap de

Creus canyon has recognized a large degree of inter-annual variability in DSWC, as well as its complex interactions with concurrent storms and down-welling transport (Ribó *et al*., 2011; Martín *et al*., 2013; Rumín-Caparrós *et al*., 2013).

3.9. Internal waves and solitons

Internal waves typically have much lower frequencies (periods from several tens of minutes to days) and higher amplitudes (up to hundreds of meters) than surface waves (Fig. 1). Internal waves can also propagate vertically through the water column transferring energy in either direction between shallower and deeper levels (Shanmugam, 2012b, 2013a, 2014, Fig. 12). These phenomena occur throughout the oceans, as evident in temperature, salinity and current measurements (Kantha and Clayson, 2000). They appear as surface roughness produced by interference between the current velocity and the upper layer background dynamics (Bruno *et al*., 2006).

Internal waves occur in stable, stratified fluids due to the restoring action of the buoyancy forces on water parcels displaced from their equilibrium position (Fig. 12). Interfacial waves on the density interfaces offer a good example of internal waves (Gill, 1982). When the density interface is disturbed, waves radiate horizontally along the interface if the vertical density gradient is high enough (Shanmugam, 2012a, 2013a, b). In continuously stratified waters internal waves travel along paths (wave characteristics) that run oblique to the horizontal plane. The exact angle is a function of the frequency of the internal wave, the buoyancy or Brunt Väisälä frequency N and the Coriolis parameter f (Gill, 1982). Perturbation of a region with high N values, such as steepened segments of the pycnocline, tend to travel rather horizontally along the pycnocline (Shanmugam, 2012a, 2013a, b), whereas if the vertical density gradient is more uniform the internal perturbation may radiate at an angle relative to the vertical (Baines, 1982). The frequencies of the free internal waves are confined to frequencies between f and N. In the upper ocean at middle and high latitudes, N typically exceeds f by one or two orders of magnitude. At low frequencies approaching f, Earth's rotational effects are critical and the waves are referred to as near-inertial internal waves (Garrett, 2001; Puig *et al*., 2001; van Haren *et al*., 2013). At higher frequencies approaching N, rotational effects become negligible. In the deep oceans at middle and high latitudes, f and N are comparable. The weak stratification at the abyssal regions prevents the formation of high-frequency internal waves (e.g., only low frequency close to f stand). In latitudes approaching the equator, f approaches zero. The diminished rotational component affects the internal wave structure and dynamics primarily at very low frequencies.

As mentioned above internal waves occur due to the disruption of horizontal density surfaces within a stratified water column. The most usual perturbing mechanism is the interaction between currents and the bottom topography (Fig. 12) in regions where it changes more or less abruptly, as ridges, banks, slopes, shelf breaks, etc. (Farmer and Armi, 1999; Shanmugam, 2013a, b). Barotropic tidal currents are often behind this type of forcing. Other forces that generate vertical perturbations also produce internal waves by the same mechanism. These types of forces may derive from the wind stress on the sea surface or from currents induced by other types of barotropic waves, such as continental shelf waves or edge waves (Bruno *et al*., 2006).

The topographic perturbations often start with a single wave or soliton, which disintegrates in rank, ordered internal wave trains (Vlasenko *et al*, 2009) of lesser wavelengths and shorter periods, relative to the original internal wave that approached the pycnocline (Apel *et al*., 1995). These secondary perturbations exhibit relatively high amplitudes and have induced currents that reach intensities of 200 cm/s.

Internal waves have been described in coastal areas of California (Emery, 1956), in the Sea of Japan (Navrotsky *et al*., 2004), the Indian Ocean (Santek and Winguth, 2005), the Bay of Biscay (Baines, 1982); Messina Strait (Brandt *et al.*, 1997); Gibraltar Strait (Brandt *et al.*, 1996; Bruno *et al.*, 2006) and on other areas along the Iberian margins (Puig *et al*., 2004; Hernández-Molina *et al.*, 2011, 2014; Van Haren *et al*., 2013; Ribó et al., 2015a).

The energy associated with internal waves is particularly important close to the continental margins (maximum horizontal velocities up to 200 cm/s and vertical velocities of 20 cm/s, Shanmugam, 2012a, 2012b, 2013a). Internal waves may act as the primary mechanism in along-slope and across-slope processes, maintenance of intermediate and bottom nepheloid layers (McCave, 1986; Dickson and McCave, 1986; Cacchione *et al*., 2002; Puig *et al*., 2004) and in the erosion of contourite terraces (Hernández-Molina *et al*., 2009; Preu *et al*., 2013).

High amplitude internal wave generation in the Strait of Gibraltar strongly affects the Iberian margins. These processes primarily occur around the Camarinal Sill and other minor sills nearby where the tide-topography interaction generates internal bores that evolves into the well-shaped wave trains mentioned above (Armi and Farmer, 1988; Farmer and Armi, 1988; Vlasenko *et al*., 2009; Sanchez-Garrido *et al*, 2011. See also Fig. 13), with amplitudes of 50 to 100 m and wavelengths of 1 to 4 km (Armi and Farmer, 1988; Farmer and Armi, 1988; Brandt *et al*., 1996; Jackson, 2004). The wave trains extend at least 200 km into the Western Mediterranean and persist for more than 2 days before dissipating (Apel, 2000; Jackson, 2004) (Fig. 13). Solitons and internal waves have been observed in many other areas along the Iberian margins (Fig. 13), including the Gulf of Cádiz, the west coast of Portugal and the Galician and Cantabrian margins (Pingree *et al*., 1986; Pingree and New, 1989, 1991; Correia, 2003; Apel, 2004; Jackson, 2004; Azevedo *et al*., 2006; Pichon *et al*., 2013); at the NW flank of Le Danois at the Cantabrian margin (Gonzalez-Pola *et al*. 2012); and the Gulf of Valencia continental slope (van Haren *et al*., 2013; Ribó et al., 2015a), where several large fields of sediment waves intrepreted as being generated by the interaction of internal waves with the seafloor develop (Ribó et al., 2015b, c). Solitons also frequently occur along the shelf break of the Iberian margins (Apel, 2004). Although present knowledge about internal waves and solitons are limited, many features along the Iberian slope could be related to their effect, as sedimentary waves on the Galician continental margin (Fig. 14, Hanebuth *et al*., 2012; 2015). Internal waves amplify its influence within submarine canyons (e.g., Shanmugam, 2012a, 2013a, b; Puig *et al*., 2014), contributing to higher velocity bottom currents within the canyon (Quaresma *et al*., 2007; Sánchez *et al*., 2014) and reworking sandy deposits, as in the Gaviera Canyon, Cantabrian margin (Fig. 15, Gómez-Ballesteros *et al*., 2014). Sedimentary waves are very common at certain depths within submarine canyons, and some of them are reworked sedimentary waves as those described in the Setubal, Nazaret and Cascais Canyons (Arzola *et al*., 2008; Lastras *et al.*, 2009) and interpreted as being due to the action of turbiditic processes (Fig. S2). To what extent these sedimentary waves could be related to the action of internal waves on submarine canyons is unknown. The fact is that the role of internal waves on submarine canyons, gullies, etc., has been overlooked and should be considered in future multidisciplinary works.

3.10. Other-related traction current

Other processes could generate traction currents in deep-marine settings (Fig. 1), such as the tsunami-, rogue- and cyclone-related currents (Shanmugam, 2006, 2012a). Tsunami waves carry energy through the water, but do not displace the water horizontally, nor do they transport sediment. During the transformation stage, however, tsunami waves erode shallow water zones and incorporate sediment into the incoming wave. Tsunami-related traction currents can thus transport large concentrations of sediment in suspension (Abrantes *et al*., 2008; Shanmugam, 2006, 2012a). Tsunamis are also important mechanisms for triggering sediment failures, as the advancing tsunami wave front can generate large hydrodynamic pressures on the seafloor that overwhelm slope stability factors (Wright and Rathje, 2003). Similar to tsunamis, rogue waves (also known as freak waves, killer waves, monster waves, extreme events, abnormal waves, etc.) and cyclone waves are categorized as intermittent processes, which can trigger bottom currents and generate large hydrodynamic pressures on the seafloor that produce submarine mudflows and slope instabilities (Shanmugam, 2012a). As such, they accelerate deep-water sedimentation and/or rework previous deposits.

The southern Iberian margins are seismically active due to the convergent plate boundary between Eurasia (Iberian sub-plate) and Africa (Nubian sub-plate) (Zitellini *et al*., 2009). Earthquakes have triggered a number of tsunamis in the historic past (Rodríguez-Vidal *et al*., 2011). These include the 1755 Lisbon tsunami (Baptista *et al*., 2003), the 1856 and 2003 tsunamis in Algeria (Roger and Hébert, 2008; Sahal *et al*., 2009) and the 1969 tsunami in the Atlantic (Guesmia *et al*., 1998). The earthquake-induced Lisbon tsunami appears in the geologic record from Portugal to the Gibraltar Strait (i.e., Dawson *et al*., 1991, 1995; Hindson *et al*., 1996; Dabrio *et al*., 1998, 2000; Hindson and Andrade, 1999; Luque *et al*., 1999, 2001, 2004; Luque, 2002; Whelan and Kelletat, 2005; Viana-Baptista *et al*., 2006; Gracia *et al*., 2006; Cuven *et al*., 2013) and even in southern England (Banerjee *et al*., 2001). The geologic record also shows evidence of ancient tsunamis impacting southwestern Spain, Portugal, Morocco (Galbis, 1932; Campos, 1991; Reicherter, 2001; Luque *et al*., 2001, 2002; Luque, 2002; Ruiz *et al*., 2004, 2005; Scheffers and Kelletat, 2005), and the Alboran Sea (IGN, 1991; Reicherter and Becker-Heidmann, 2009).

Few works have considered the deep-sea sedimentary implications of tsunamis that have impacted the Iberian margins. The 1522 Almeria earthquake (M>6.5) affected a large area of the western Mediterranean, even though its epicenter occurred in the Gulf of Almeria along the Carboneras Fault Zone. The earthquake triggered submarine slides in the Gulf of Almeria and tsunami waves (Reicherter and Becker-Heidmann, 2009). The IODP Expedition 339 identified sandy deposits along the Gulf of Cádiz and west of the Portuguese slope (Fig. 16) as possible tsunami deposits (Stow *et al*., 2013b). Other tsunami-related traction currents and deposits may occur along the Iberian slopes. While evidence of other events may exist in the study region, researchers have not established consistent sedimentological criteria for identifying paleo-tsunami deposits (Shanmugam, 2012b). Deeper water tsunami deposits and structures may resemble their shallow water counterparts, or might resemble turbidity and debris (or mud) flow sequences in terms of facies associations (Dawson and Stewart, 2007).

4. Morphologic and sedimentary implications

The descriptions above highlight a number of permanent and intermittent processes that affect bottom currents. These processes could be very variable as overflows, tides, eddies, deep-sea storms, secondary circulation, internal waves and tsunamis, rogue waves and cyclone waves. All these processes show some degree of variability and, although one of them can prevail in a particular situation, more probably there are several processes acting in combination to determine the bottom current’s local direction and velocity. Many of these processes and their effects on deep-water sedimentation are not well understood. Along the Iberian margins, several of these processes interact locally or regionally with the seafloor to affect its morphology and sediment distribution (Table-I). We cannot however yet discern specific sequences of events or the relative influence of these processes from sedimentary evidence. Discussions and debate continue on how bottom currents form the range of morphologies observed among contourites. To stimulate this discussion, future analyses must include the wider and more complex range of deep-water processes described above. While local oceanographic setting provides first order constrains on contourite morphology, ancient contourite examples from the geologic record can also reveal how they may vary in response to different deep-marine conditions.

Specific bottom current facies have been described by many authors (see a recent compilation in Rebesco *et al*., 2014). The standard contourite facies model bi-gradational sequence was first proposed by Gonthier *et al*. (1984) and Faugères *et al*. (1984), based on the Faro Drift within the middle slope of the Gulf of Cádiz. This general model for contourites represents the transition from weaker to stronger bottom current flow, and then back to weaker conditions (Stow and Holbrook, 1984; Stow *et al*., 2002a; Huneke and Stow, 2008). The bi-gradational facies sequences are observed among ancient to recent contourite deposits, including the Brazilian margin (Viana and Faugères, 1998), the Irish margin (Øvrebø *et al*., 2006) and in ancient deposits (e.g., Chinese sedimentary basins, Gao *et al*., 1998). Authors commonly report partial or incomplete contourite sequences (Howe *et al*., 1994; Stoker *et al*., 1998; Shanmugam, 2000; Howe *et al*., 2002; Stow *et al*., 2002a, 2013a; Mulder *et al*., 2013). Further modifications of the contourite model (e.g., Stow *et al*., 2002a, b; Stow, 2005; Stow and Faugères, 2008) have demonstrated a considerable degree of variation, thus complicating the establishment of a unique and systematic facies designation. Mulder *et al*. (2013) have recently demonstrated that a particular facies sequence may depend on the supply of coarser terrigenous clastic material provided by increased erosion of indurated mud along the flanks of confined contourite channels (mud clasts), by increased sediment supply from rivers (quartz grains) and/or by downslope mass transport along the continental shelf and upper slope. These results are consistent with the findings of Masson *et al*. (2010), that the classical contourite depositional sequence proposed by Gonthier *et al*. (1984) should be interpreted with special consideration given to the regional sedimentological background.

The Faro Drift deposits do not offer the best example of a contourite facies sequence due to the preponderance of mud within these deposits, and their location along the distal part of a much larger contourite depositional system (CDS). Other facies from different parts of the same depositional system do not fit neatly into the conceptual model suggested by the Faro Drift (e.g., Mulder *et al*., 2013; Stow *et al*., 2013a; Hernández-Molina *et al*., 2013, 2014). Mulder *et al*. (2013) have in fact showed that most of the contacts between the classical contourite facies (mottled, fine sand, coarse sand) are rather sharp than transitional, an observation that supports the interpretation by Shanmugam (2006, 2012a, 2013b). The proposed facies sequence for the Faro Drift could therefore serve as an appropriate model for fine-grained contourite deposits, for which pervasive bioturbation is a diagnostic feature. This interpretation, however, suggests that the Faro Drift does not offer a representative example of other types of contourite deposits (Martín-Chivelet *et al*., 2008; Shanmugam, 2012a, 2013b; Mulder *et al*., 2013). Other authors have described contourite settings with a higher proportion of sand (Shanmugam *et al*., 1993; Shanmugam, 2000, 2012a, 2013b), indicating that activity of bottom currents prevails over bioturbation.

This controversy in contourite identification demonstrates that contourites exhibit greater variation than the established facies model suggests and commonly include traction sedimentary structures. The contourite facies model thus would benefit from further modification. A consistent facies model however faces substantial challenges in terms of the wide range of oceanographic permanent and intermittent processes (and their spatial and temporal variation) that can influence development of CDS.

We began an effort to reassess of contourite facies during the IODP Expedition 339 along the Iberian margins, in the Gulf of Cádiz and off western Portugal (http://iodp.tamu.edu/scienceops/expeditions/mediterranean_outflow.html), (Expedition 339 Scientists, 2012; Stow *et al*., 2013b; Hernández-Molina *et al*., 2013). The facies included sand-rich, muddy sand, silty-mud and mud-rich contourites, all of which formed at moderate (20-30 cm/ky) to very high (> 100 cm/ky) rates of sedimentation. The contourites recovered during this expedition were remarkably uniform in composition and textural attributes. The muddy and silty contourite deposits displayed an absence of primary sedimentary structures, but exhibited intense continuous bioturbation throughout the section sampled. The cores also showed consistent, bi-gradational sequences with inverse and then normal grading as well as a number of partial sequence types. Expedition 339 also identified thick and extensive contourite sand deposits, as well as turbidite sands apparently reworked by bottom currents (Expedition 339 Scientists, 2012; Stow *et al*., 2013b; Hernández-Molina *et al*., 2013). These sands occurred in the proximal part of the CDS close to the Strait of Gibraltar (Fig. 17), where Expedition 339 retrieved a very thick and sandy contourite layer (> 10 m) that showed traction sedimentary structures (Hernández-Molina *et al*., 2013, 2014; Brackenridge, 2014). Sand-rich contourites also occur in other areas around Iberia, such as the Galician and Cantabrian margins (Fig. 17) (Alejo *et al.*, 2012). Future work around the Iberian margins will systematically categorize these sedimentary facies and frame them according to bottom current dynamics and other associated oceanographic processes (e.g., overflows, barotropic tidal currents) including intermittent processes (e.g., vertical eddies, deep sea storms, horizontal vortices, internal waves and tsunamis).

Sedimentologists have debated the differentiation between contourites and turbidites for almost five decades (Hollister, 1967; Piper, 1972, Hollister and Heezen, 1972; Bouma, 1972, 1973; Bouma and Hollister, 1973) but have not definitively established which structures distinguish contourites. Contourite processes can in fact trigger gravitational collapse that forms submarine lobes as in the Gulf of Cádiz (Habgood *et al*., 2003; Hanquiez *et al*., 2010) or rework previous turbiditic deposits (Shanmugam *et al*., 1993, Shanmugam 2012a, 2013b). Contourite and turbiditic processes can also intermingle in both vertical and lateral dimensions, and thus form mixed deposits. Differentiating contourite from turbidite represents a major challenge for future research along the Iberian margins, especially given the legacy of emphasis on mass-transport and turbiditic processes in interpretations of Iberian margins deposits.

Regional interpretations of the Iberian margins have generally underestimated the role of bottom currents in shaping the seafloor and controlling the sedimentary stacking pattern. Very large contourite drifts occur along these margins (see compilation at Llave *et al*., 2015a, b) and small to intermediate sized drifts are ubiquitous (e.g., Ercilla *et al*., 2015). While CDS genetic models are under revision, bottom currents and associated oceanographic processes clearly control the physiography and sedimentation around the Iberian continental margins and adjacent basins. From all aforementioned oceanographic processes associated to bottom currents, the position of interfaces between major water masses and their vertical and spatial variation in time specifically appears to exert primary control in determining major morphologic changes (changes along the slope gradient), including the formation of the contourite terraces (Fig. 4, 5 and 18). Density contrasts that form boundaries (Hernández-Molina *et al*., 2011; Ercilla *et al*., 2015; Llave *et al.*, 2015a, b) similar to those reported for other margins (e.g., Hernández-Molina *et al*., 2009; Preu *et al*., 2013) are important components of these interfaces due to their responses to internal waves, tidal waves, deep-sea storms, eddies, secondary circulations and other processes. A consistent CDS model thus requires monitoring of these processes and an accurate cause and effect description of how these processes affect deep-water sedimentation, sedimentary structures and CDS architecture.

5. Final considerations

5.1. A new multidisciplinary approach

Bottom current-controlled depositional, erosional and mixed morphological features commonly occur along the Iberian continental margins at different depths and in various settings. This paper depicts the bottom current circulation around the Iberian continental margins that are affected by a number of permanent and intermittent oceanographic processes (Fig. 19). Previous models for the margins have not fully acknowledged the role of these processes and they continue to elude our fullest understanding. Internal waves, tidal waves, deep-sea storms, eddies, secondary circulations and other deep-sea processes generally shape the seafloor over short- and/or long-term time scales. Future research should seek to interpret these oceanographic processes in sedimentary stacking patterns and the overall architecture of continental margin deposits.

A clearer understanding of oceanographic processes related to bottom currents and their associated sedimentary features requires a multidisciplinary approach, given the complexity of the phenomena under consideration. This approach should integrate frameworks of marine geology, physical oceanography and benthic biology. A combined analysis of sedimentology and fluid dynamics is a priority objective. The approach should specifically include next topics:

- Detailed studies of deep-water oceanographic processes related to flow phenomena, including internal waves, tides, benthic storms, eddies and vortices with the objective of understanding their importance, interactions and influence on bathymetric features.
- A more systematic understanding of the depositional, erosional and mixed morphological features, as well as their evolution over time and distribution at different points along the margin. Linking these features to specific oceanographic processes will help categorize small and large-scale features and clarify sedimentary facies associations.
- Studies about the general physiographic architecture of the margins and its relation with water masses structure and associated interphases require more attention.
- Numerical modelling can help link the variation in genetic processes to the range of observed erosional and depositional features.
- Detection and characterization of past pycnoclines based on proxies from the sedimentary record represents an important objective for future multidisciplinary research. Estimation of the depth range of water mass interfaces and the study of processes associated with these interfaces (such as baroclinic tides and internal waves) can inform the interpretation of contourite terraces along the Iberian continental slope. Paleoceanographic models can help determine potential periods of greater density contrasts within the pycnocline. Contourite terraces suggest relict features along the slope that hint at colder oceanic conditions in the past. These hypothetic conditions require further investigation due to their implications for regional and global climate dynamics.
- Research should also seek to detect and characterise intermediate and deep nepheloid layers. These layers often occur at water mass interfaces. Benthic boundary layers are also critical interfaces that determine seafloor morphology. These interfaces have mostly been studied from a biogeochemical perspective and have not been systematically integrated in the sedimentological and oceanographic models.
- We require a more comprehensive understanding of fluid dynamics around submarine obstacles.
- Oceanographic monitoring should also focus on linking erosional contourite morphological features to specific bottom current conditions (e.g., velocity, energy).
- Analysis of sedimentary evidence and ROV-assisted visual observations should relate modern erosional, depositional and mixed features to bottom current and associated processes. This analysis should query local outcrops on the seafloor, including channel flanks, slide scarps, etc., and synthesize observation with facies associations evident from cores (IODP, etc.) and other sedimentary evidence.

- A wide range of small-scale features (cores, outcrops and their lithologic facies), and large-scale features (depositional systems and seismic facies) require more specific linkages to oceanographic processes.
- Research should foster the analysis and evaluation of the influence of oceanographic processes on fragile geohabits, relict or active today, which could be situated in scenarios of risk because of deep environmental changes derived from water-mass circulation modifications.
- The current facies model must be adapted to include sandy facies. The facies model should also address interactions between bottom current and turbidite processes, and offer consistent interpretations of both sandy contourites and bottom-current reworked turbidite sands that occur along the Iberian margins.

The differing objectives, frameworks and terminology used in marine geology, physical oceanography and benthic biology make these integrative efforts a formidable task. Some of the challenges can be circumvented by deploying remote sensing technology (e.g., Automated Underwater Vehicles, e.g., AUVs and Remote Operated Vehicles, ROVs) and capitalizing on advances in data acquisition and processing to build a more complete picture of water column dynamics, seafloor morphology, contourite distribution and benthic community structure. Further development of seismic oceanographic techniques (Buffett *et al*., 2009; Carniel *et al*., 2012; Pinheiro *et al*., 2010) can help integrate *in-situ* observations with 2D (or 3D) images of the seafloor and water column (Fig. 18) while acoustic approaches integrated with oceanographic data such as CTD or (L) ADCP (Preu *et al*., 2011, 2013; Hernández-Molina *et al*., 2014) and instrumented moored stations (Rebesco *et al*., 2013). High-resolution, 3D seismic images (Campbell and Deptuck, 2012), multibeam and backscatter data (Palomino *et al*., 2011; Sweeney *et al*., 2012) will also help characterize contourite features and correlate them with specific oceanographic processes.

5.2. Applied research

Multidisciplinary approaches to understanding contourite development carry major implications for climate, resource exploration and basic science. Sandy contourites represent an entirely different deep-water sand deposit than turbidite sands. Given the prominent role of turbidite sands in deep-water oil and gas plays, understanding how their contourite counterparts form could fundamentally change deep water exploration paradigms (Viana, 2008; Stow *et al*., 2013b; Hernández-Molina *et al*., 2013). We will continue to explore the contourite processes and their implications for resource exploration, using IODP Expedition 339 data. These data show extensively distributed and well / moderated sorted sand deposits (Expedition 339 Scientists, 2012; Stow *et al*., 2013b; Hernández-Molina *et al*., 2013) that can potentially serve as reservoirs units, as well as muddy contourites that may function as hydrocarbon seals or source rocks and/or unconventional reservoirs (Viana, 2008;

Shanmugam, 2012a, 2013a, b; Brackenridge *et al*., 2013; Stow *et al*., 2013b). The frequent association of both sandy and muddy contourite deposits with cold-water coral mounds (Huvenne *et al*., 2009; Van Rooij *et al*., 2011; Somoza *et al*., 2014; Sánchez *et al*., 2014), which may function as unconventional reservoirs (Henriet *et al*., 2014) further demonstrate the potential for plays in these settings. As the climate warms due to the usage of these fossil fuels, we must also consider the role of deep-water circulation as a climate modulator. In addition to its effects on global circulation, variability in deep-water circulation also affects deep-water ecosystems, such as reefs. Growing interest in seafloor mining of metalliferous resources (Hoffert, 2008; Rona, 2008) will also benefit from better understanding of bottom current and associated processes dynamics. These types of deposits have been identified along the Iberian margins (e.g., González *et al*, 2009, 2010a, b). Mapping of these deposits can help interpret their formation with respect to bottom currents and seafloor morphology parameters.

6. Conclusions

Based on this review work achieved by Geologists, Physical Oceanographers, Paleoceanographers and Benthic Biologists, we highlight that bottom current dynamics are commonly affected by a number of associated, permanent and intermittent, oceanographic processes, as overflows, tides, eddies, deep-sea storms, secondary circulation, internal waves and other -related traction currents (e.g., tsunamis). All these processes show some degree of variability and, although some of them can prevail in a particular situation, it is more probably to find several processes acting in combination determining the bottom current's local direction and velocity. Many of these processes and their effects on deep-water sedimentation are not understood. Along the Iberian margins, several of these processes interact locally or regionally with the seafloor affecting its morphology and sediment distribution, and three major conclusions could be considered:

a) Contourite depositional and erosional features are ubiquitous along the margins, indicating that bottom currents and associated oceanographic processes clearly control the physiography and sedimentation around the Iberian continental margins.

b) The position of interfaces between major water masses and their vertical and spatial variations in time specifically appears to exert primary control in determining major morphologic changes along the slope gradient, including the contourite terraces development. Density contrasts that form boundaries are important components of these interfaces due to their responses to internal waves, tidal waves, deep-sea storms, eddies, secondary circulations and other processes.

c) Contourite deposits exhibit greater variations than the established facies model for these deposits suggests. Therefore, a consistent facies model however faces substantial challenges in terms of the wide range of oceanographic permanent and intermittent processes that can influence in their development.

Future models thus require the monitoring of these oceanographic processes along the Iberian continental margins (as in another margins) and an accurate cause/effect description of them in terms of deep-water sedimentation, sedimentary structures and architecture. This approach should integrate frameworks of marine geology, physical oceanography and benthic biology.

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FIGURE CAPTIONS

Figure 1. 3D sketch depicting the possible oceanographic processes in deep-water environments. In addition to density currents and overflows, the velocity at the seafloor can also be affected by barotropic currents or intermittent processes like cascading, giant eddies, deep sea storms, vortices, internal waves, internal tides, tsunamis, cyclone waves and rogue waves (From Rebesco *et al*., 2014, *with permission from Elsevier*).

Figure 2. Swath bathymetry at the exit of the Strait of Gibraltar. Main depositional and erosive features are shown (from Hernández-Molina *et al*., 2014 *with permission from GSA*). Note that furrows are oblique to the main channels due to the secondary circulatipon related to the benthic Ekman transport.

Figure 3. Main morphologic features around the Theta Gap showing the regional water mass circulation.

Figure 4. Example of three contourite terraces (Ortegal, Pardo Bazán and Castro) along the continental slope northwest of Galicia and its relation with the interfaces between water masses. A) Multibeam bathymetry indicating the position of the terraces at water depths around 0.2-0.4 km; 0.9-2 km and 2.2-2.4 km (adapted from Maestro *et al*., 2012); B) Seismic profile (CO-16) acquired during the CONTOURIBER-2 Cruise crossing the terraces (line position in A). Interfaces between regional water masses at present time is included, with the theoretical position of MOW during glacial stages (in blue); C) T/S diagram corresponding to eight profiles from ARGOS buoy (6900469, PROVOR Profaling Float CTS3- OVIDE Project) between the 9.11.2012 to 18.01.2013 (from Sánchez-González, 2013); D) Water masses percentage in the adjacent area to the contourite terraces (from Sánchez-González, 2013).

Figure 5. A) Bathymetric map of the west Iberian margin, showing the trajectories of the RAFOS floats that were caught by Meddy-9 (blue line) and the cyclone C (magenta line) reported from Pinheiro *et al*. (2010); the yellow fractions of these trajectories correspond to the period between the 28th of August and the 3rd of September 1993, during which the seismic profile in B and C was acquired. B and C) Complete migrated seismic section showing the eddies (meddies) related to the Mediterranean Outflow Water (MOW) and their interaction with the seafloor (Pinheiro *et al*., 2010), including the water masses and the interaction of MOW with a contourite terrace in the middle slope. The segment to the East of the line was acquired from W to E, from the 2nd till the 3rd of September 1993. Approximate depth based on 1500 m/s sound speed, and longitude axis are indicated; C) Detail of a migrated

section with interpreted upper (MU) and lower (ML) MOW (Pinheiro *et al*., 2010); and D) detail of a migrated section with the interpreted upper MU (Pinheiro *et al*., 2010). High amplitude reflections within the profiles, especially to the upper and lower boundaries of MOW and meddies are indicative of density contrast surfaces in the water column.

Figure 6. Maps showing: A) the M2 barotropic current velocity map. The maximum M2 velocities (cm/s) are illustrated by the tidal ellipse semi-major axis magnitude; B) K1 barotropic current velocity phase map. This corresponds to the phase lag of the maximum current behind the maximum tidal potential of M2 (degrees referenced to GMT); C) Semi-diurnal barotropic forcing term map (1/s). Internal tide generation **"**hotspots**"** are pointed out by the following symbols: AC= Arosa canyon; AV= Aveiro canyon; CB= Conil-Barbate; CF= cape of Finisterra; CS= cape of Sagres; CO= cape of Ortegal; EC= Estremadura promontory; GB= Galician bank; GS= Gettysburg seamount); MC= Murgia Canyon; NC= Nazaré canyon; OP= Ortegal promontory; OS= Ormonde seamount; PC= Porto canyon; TP= Tagus Plateau; Vc= S. Vicente canyon (from Quaresma and Pichon 2013).

Figure 7. Examples from sandy deposits within the Cadiz Contourite Channel affected by bottom current and barotropic and internal tides (probably generated at the continental slope), which reinforce the normal MOW flow (from Stow *et al*., 2013a). A) Cadiz Contourite Channel image derived from swath bathymetric data, showing location of bottom photograph transects (black) taken with a BENTHOS-372 camera; B) Airgun deep monochannel seismic profile across the Cadiz Contourite Channel (location in A); C, D and E) Bottom photographs along the Cadiz Contourite Channel: Deep linear scours (or small furrows) on the sea bottom with very rare starved ripples of coarse sediments (C); Strait to rhomboidal asymmetrical starved ripples; linear to sinuous asymmetrical ripples (D); Sinuous to crescent sand waves, observed in the Cadiz secondary channel (E). Stoss side with small regular asymmetrical ripples, lee side with linear erosional chutes, and sand wave trough region with more irregular ripple pattern. Both ripple and dune orientation indicates flow towards the west, whereas the current vane shows more southward flow direction at the time of the photo. Compass diameter 7cm; current vane 25cm; F) Bottom current directions as inferred from bedform measurement for Transects 2 and 5 (left) and bottom current directions during the running of Transect 4 as inferred from current vane orientation (right); G) Tidal data for 16th and 17th September 2001 (Port of Cadiz), together with an indication of the duration of each photographic transect (up), and tidal amplitude through the month of September 2001 (down).

Figure 8. Time series observations of the near-bottom current (78 m above the seafloor) at the Exit of the Strait of Gibraltar (A) for 79 days at 472 m water depth and at the eastern Gulf of Cadiz (B) for 97 days at 554 m water depth (see location in Figure 2), showing a strongly directional MOW towards the WSW (A) and NNW (B) with speeds up to 1 m/s. A detail in (A) is illustrating the relation between the current speeds and temperature from 10-14 September of 1975. Inset in B depicts a semi-diurnal tidal periodicity with a spring-neap tidal cycle is clearly observed, which also modulates the near-bottom temperature. The jump in the temperature record in B is presumable due to the mooring turn-around operation and the sampling of a slightly different water parcel after its redeployment. *Published with permission from British Oceanographic Data Center (BODC) and National Oceanographic Centre, Southampton (NOCS).*

Figure 9. Time series observations of near-bottom salinity, potential temperature, current speed and suspended sediment concentration (SSC) recorded at the Catalan continental rise (1890 m in depth), illustrating an example of a benthic storm occurring in the northwestern Mediterranean basin after the winter 2009 deep dense water formation event. Note the sharp increase of salinity and temperature associated to the arrival of the dense waters to the basin and the subsequent increase of near-bottom velocities with a peak >40 cm/s that caused resuspension of bottom sediments (modified from Puig *et al*., 2012).

Figure 10. A) A snapshot of simulated zonal component of velocity and salinity at 7.5°W showing the MOW undercurrents flowing westwards in the Gulf of Cadiz; B) Snapshot of simulated salinity and zonal component of velocity at 13°W, showing the large vertical extension of the velocity signal associated with the MOW eddies (meddies) downstream of Portimao Canyon (from Serra *et al*., 2010a).

Figure 11. A) Map of the Gulf of Lions area showing the regions of dense water formation and distribution (adapted from Puig *et al*., 2013). B) A dense shelf water cascading (DSWC) event in the Gulf of Lions' shelf illustrated by a time series of near-bottom potential temperature, current speed and suspended sediment concentration (SSC) recorded at 500 m depth (5 m above bottom) in the axis of Cap de Creus Canyon during the winter 2004-05. Adapted from Puig *et al.* (2008). Legend for water masses by alphabetic order: AW= Atlantic Water; LIW= Levantine Intermediate Water; WIW= Western Mediterranean Intermediate Water; WMDW= Western Mediterranean Deep Water.

Figure 12. A) Schematic diagram showing the position of the pycnocline (i.e., primary density stratification), where density gradient is the sharpest, between mixed (upper) and deep (lower) ocean layers of different densities (adapted from Shanmugam, 2013b). Internal

waves and tides propagate along the boundaries of both primary and secondary density stratifications. Internal waves and astronomical internal tides propagate along the boundaries of density stratification in deep-marine environments. Barotropic currents (red arrow) are generated by surface waves and tides, whereas baroclinic currents (dark blue arrow) are generated by internal waves and tides; B) Conceptual oceanographic and sedimentologic framework showing deposition from baroclinic currents on continental slopes, in submarine canyons, and on guyots. On continental slopes and in submarine canyons, deposition occurs in three progressive stages: (1) incoming internal wave and tide stage, (2) shoaling transformation stage, and (3) sediment transport and deposition stage. Continental slopes and submarine canyons are considered to be environments with high potential for deposition from baroclinic currents. In the open ocean, baroclinic currents can rework sediments on flat tops of towering guyot terraces, without the need for the three stages required for the deposition on continental slopes (From Shanmugam, 2012, 2013b).

Figure 13. Selected examples of internal solitons propagating toward the continental slope and shelf around the Iberian margins from: A) Catalonian margin; B) East of the Strait of Gibraltar; C) West of the Strait of Gibraltar; D) Cádiz; E) NW Morocco; F) NW Portugal; G) Galicia Bank; H) Rias Baixas (Galicia margin); I) Cantabrian margin (off Gijon); J) Cantabrian margin (off Bilbao). These oceanographic processes are ubiquitous around Iberia as well as in the adjacent continental margins from Morocco, France and Porcupine related mainly to the shelf break, seamounts and banks. Their real implications in sedimentary processes have not been considered until now although they represent energetic and constant oceanographic processes. *Images are synthetic aperture radar images [from Apel, 2004 and EOLi (Earth Observation Link, European Space Agency, http://earth.esa.int/EOLi/EOLi.html)]*.

Figure 14. Example of sedimentary waves in the Galician continental margin, which could be attributed to internal waves. A) Swath bathymetric map of the margin indicating the position of the sedimentary waves along the NW flank of a contourite drift in the lower slope; B) Seismic profile across the sediment waves (*data courtesy from the MARUM project GALIOMAR*, from Hanebuth, *et al*., 2012).

Figure 15. Examples of sandy bedforms related to internal tides and waves within the Gaviera Canyon, in the Cantabrian margin. A) Regional area location; B) Shaded bathymetric map of part of the Cantabrian Sea margin (isobaths equidistance 100 m) indicating the position of the major canyons, with illumination from the NW (Gómez-Ballesteros *et al*., 2013). C) Topas profile crossing NE the head of the Gaviera canyon (Gómez-Ballesteros *et al*., 2013).

Submarine photographs of the Gaviera canyon seafloor; D) at 863 m water depth (-5.78628 / 43.92902) taken the 24 April 2010 at 15:05. Water temperature was about 9.93ºC; salinity of 35.76 and potential density of 27.54 kg/m^3; E) at 837.6 m water depth (-5. 77847°E / 43. 93003°N) taken the 26 April 2010 at 15:41. Water temperature was about 9.95°C; salinity of 35.75 and potential density of 27.53 kg/m^3 (*Photo taken by F. Sánchez, IEO*).

Figure 16. A) Regional water masses and Gulf of Cádiz CDS site locations sampled during IODP Expedition 339; B) Sandy Holocene deposits drilled during IODP Expedition 339 (Sites U1388 and U1389) interpreted as tsunami deposits (Stow *et al*., 2013b, Pers. Comm., Yasuhiro Takashimizu, Niigata University, Japan).

Figure 17. Examples of deep water sandy deposits around the Iberian margins. Some of them are related to the action of contour currents, but other are related to other oceanographic processes as internal waves, tides, etc. Most of these deposits have not been studied yet and future work should clarify their characteristics and genesis. A) Examples of sandy contourite facies sampled during IODP Expedition 339, in the Gulf of Cadiz close to the exit of the Strait of Gibraltar (Modified from Hernández-Molina *et al*., 2014); B) Coarse sands and bioclastic gravel deposits in the slope of the Gulf of Cadiz at the exit of the Strait of Gibraltar collected with the CONTOURIBER Project (2010); C) Laminated sandy contourite facies, showing inclined lamination as a result of sand wave migration in the Gulf of Cadiz sand sheet in proximal sector (Stow and Fauguères, 2008); D) Contourite sandy deposits affected by bottom current and barotropic and internal tides in the slope channels of the middle slope of the Gulf of Cadiz (from Stow *et al*., 2013a); E) Sandy turbiditic (reworked?) facies from ISIS from ca. 4300 m water depth in the Nazare Canyon (*courtesy from V. Huvenne of NOCS, UK*); F) Sandy deposits on top of the Galicia Bank by bottom current (*INDEMARES Project, Photo taken by F. Sánchez, IEO, Spain*); G) Example of sandy deposits from the middle slope of the Ortegal Spur (*From A Selva Cruise, 2008 and R/V Belgica 09/14a cruise, EC FP7 IP HERMIONE project*); H) Submarine photographs of reworked sandy deposits within the Gaviera canyon, Cantabrian Sea (*Photo taken by F. Sánchez, IEO, Spain*); I) Sandy deposits along the SE flank of the Le Danois Bank in the slope of the Cantabrian Sea (*ECOMARG 3 Project, Photo taken by F. Sanchéz, IEO, Spain*); and J) Sandy deposits in the Porcupine seabight, Irish continental slope (*courtesy of MARUM, Germany*).

Figure 18. Examples of combination of physical oceanographic and geologic/geophysical data showing the relationship between the long-term current regime, seafloor morphology and sub-bottom sediment geometry in the Gulf of Cádiz, from the exit of the Strait of Gibraltar

(Hernández-Molina *et al*., 2014, *with permission from Geological Society of America*). A) Salinity and temperature data in the water columns. The black lines refer to isopycnals and neutral density (kg/m^3); B) Vertical sections of ocean velocity measured at the eastern area close to the exit of the Strait of Gibraltar. The water colour code displays the Acoustic Doppler Current Profiler (ADCP) velocity components. The panels on the left correspond to the velocity component in the east-west direction (positive values indicate currents towards the east), and the panels on the right correspond to the velocity component in the north-south direction (positive values indicate current towards the north). The dashed black lines indicate where the current velocity is zero. See Figure 2 for section locations.

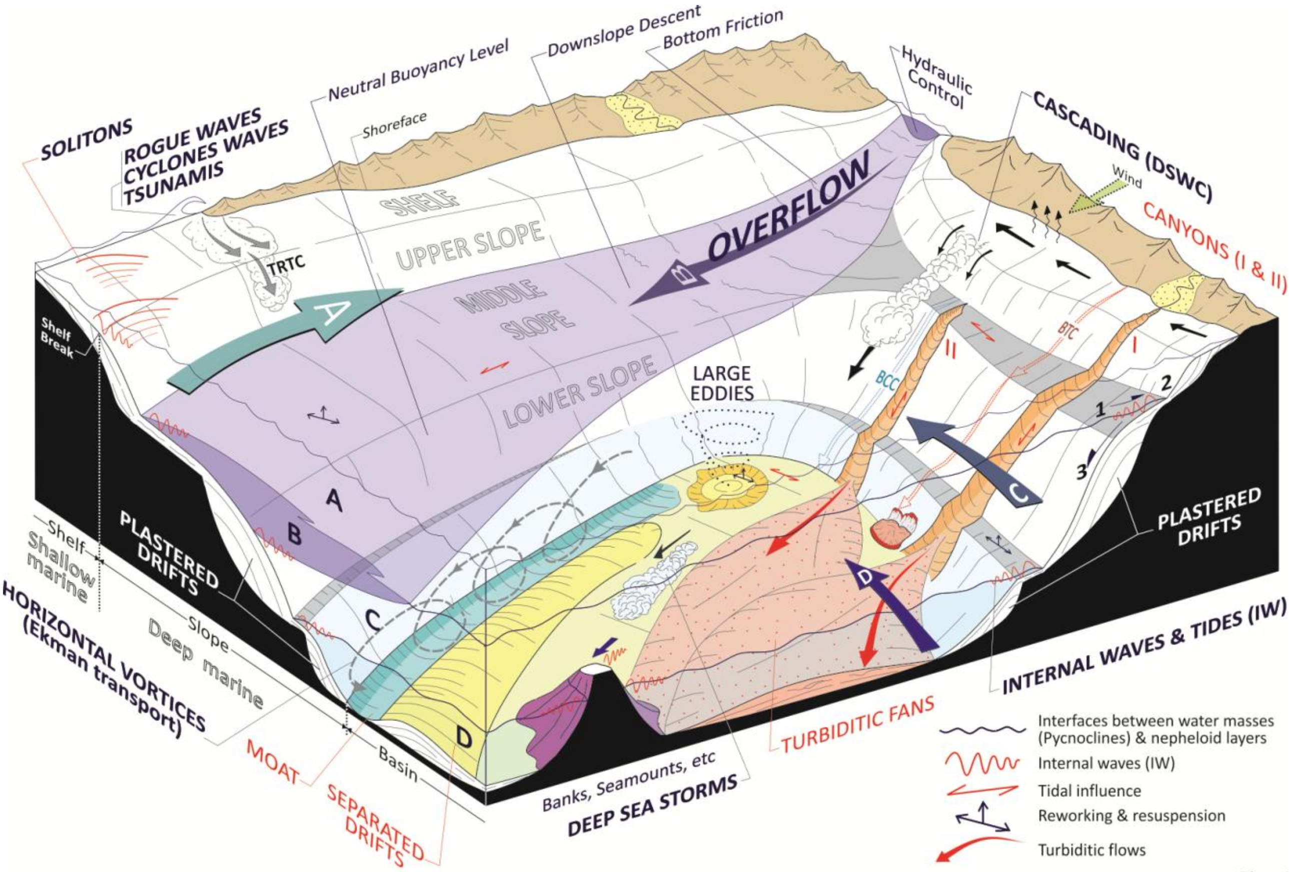
Figure 19. Sketch summarizing the locations of permanent or intermittent oceanographic processes along the Iberian margins and their short- and long-term controls on sedimentary facies and seafloor morphology. These processes include overflows, barotropic currents, tides, -eddies, deep-sea storms, horizontal vortices, internal waves, tsunamis, rogue waves and cyclonic waves. In one particular area, one of these processes may be dominant, but usually several processes act in tandem to determine the local direction and velocity of bottom currents. Legend for the Internal tide generation **"**hotspots**"**: AC= Arosa canyon; AV= Aveiro canyon; CB= Conil-Barbate; CF= cape of Finisterra; CS= cape of Sagres; CO= cape of Ortegal; EC= Estremadura promontory; GB= Galician bank; GS= Gettysburg seamount); MC= Murgia Canyon; NC= Nazaré canyon; OP= Ortegal promontory; OS= Ormonde seamount; PC= Porto canyon; TP= Tagus Plateau; VC= S. Vicente canyon. Legend for water masses by alphabetic order: AABW= Antarctic Bottom Water; ENAC= Eastern North Atlantic Current; ENACW= Eastern North Atlantic Central Water; IPC= Iberian Poleward Current; LDW= Lower Deep Water; LIW= Levantine Intermediate Water; LSW= Labrador Sea Water LSW; MAW= Modified Atlantic Water; MOW= Mediterranean Outflow Water; NADW= North Atlantic Deep Water; NASW= North Atlantic Superficial Water; SAIW= Subarctic Intermediate Water; WMDW= Western Mediterranean Deep Water.

Table-I.- Summary of the oceanographic processes and products along the Iberian margins.

SUPPLEMENTARY MATERIAL

Figure S1. A) Physical processes acting in overflows. B) Sketch of a dense overflow showing the coordinate system and some of the notations used: ambient density r, plume density r+Dr, reduced gravity g', bottom slope a, Coriolis parameter f and Nof velocity UN. Also indicated are the Ekman layer and the benthic Ekman transport. Figure modified from Rebesco et al., (2014) with permission from Elsevier.

Figure S2. Examples of sedimentary waves described in the Setúbal Canyon, west Iberian margins. A) Regional bathymetry map. Contours are every 100 m and outlined in bold every 500 m. Side scan sonar images (A, B and C) showing details of the canyon floor in the lower Setúbal Canyon; B) Transverse linear bedforms illustrating sediment facies in the Core CD56844, interpreted to comprise of coarse-grained sediment wave deposits (from Arzola et al., 2008); C) Reworked sediment waves on the canyon floor and profile show the contrast between the steep canyon walls and edge of the northern margin terrace, and the wide, relatively flat but slightly domed canyon floor (from Arzola et al., 2008); D) Sediment bedforms from the channel-lobe transition zone (from Lastras et al 2009). These sedimentary wave examples (locations in A) are described in the Setubal, Nazaré and Cascais Canyons and interpreted as being due to the actions of turbiditic processes. However, they are very common within submarine canyons worldwide and in many other cases are generated and / or reworked by oceanographic processes, as tidal current or internal waves (e.g., Shanmugam, 2012a, 2013b). The role of these processes on submarine canyons, gullies, etc. have been underestimated but should be considered in future multidisciplinary studies.



SOLITONS
ROGUE WAVES
CYCLONES WAVES
TSUNAMIS
Neutral Buoyancy Level
Shoreface
Downslope Descent
Bottom Friction
Hydraulic Control
CASCADING (DSWC)
Wind
CANYONS (I & II)
SHELF
UPPER SLOPE
OVERFLOW
MIDDLE SLOPE
LOWER SLOPE
TRTC
Shelf Break
LARGE EDDIES
BCC
BTC
PLASTERED DRIFTS
Shelf
Shallow marine
Slope
Deep marine
Basin
HORIZONTAL VORTICES (Ekman transport)
MOAT
SEPARATED DRIFTS
Banks, Seamounts, etc
DEEP SEA STORMS
TURBIDITIC FANS
INTERNAL WAVES & TIDES (IW)
Interfaces between water masses (Pycnoclines) & nepheloid layers
Internal waves (IW)
Tidal influence
Reworking & resuspension
Turbiditic flows

Fig. 1

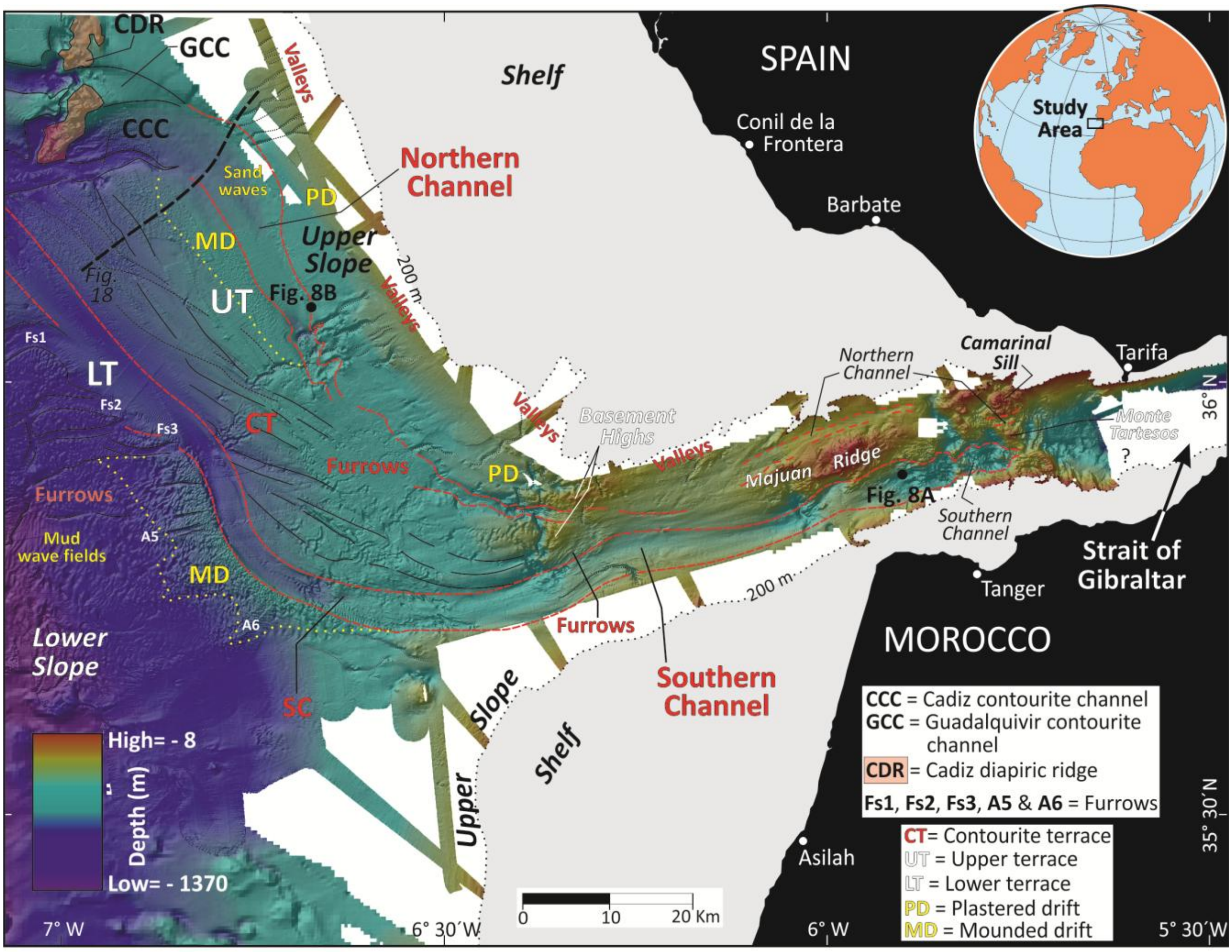

CDR
GCC
CCC
Valleys
Shelf
SPAIN
Conil de la Frontera
Barbate
Study Area
Northern Channel
Sand waves
PD
MD
Upper Slope
200 m
Fig. 18
UT
Fig. 8B
Fs1
LT
Fs2
Fs3
CT
Furrows
Basement Highs
Northern Channel
Camarinal Sill
Tarifa
36° N
Monte Tartesos
Majuan Ridge
Fig. 8A
Southern Channel
Mud wave fields
A5
A6
Strait of Gibraltar
Tanger
MOROCCO
Lower Slope
SC
Southern Channel
Slope
Shelf
Upper
High= - 8
Depth (m)
Low= - 1370
7° W
6° 30´W
0
10
20 Km
6° W
Asilah
5° 30´W
35° 30´N
CCC = Cadiz contourite channel
GCC = Guadalquivir contourite channel
CDR = Cadiz diapiric ridge
Fs1, Fs2, Fs3, A5 & A6 = Furrows
CT= Contourite terrace
UT = Upper terrace
LT = Lower terrace
PD = Plastered drift
MD = Mounded drift

Fig. 2

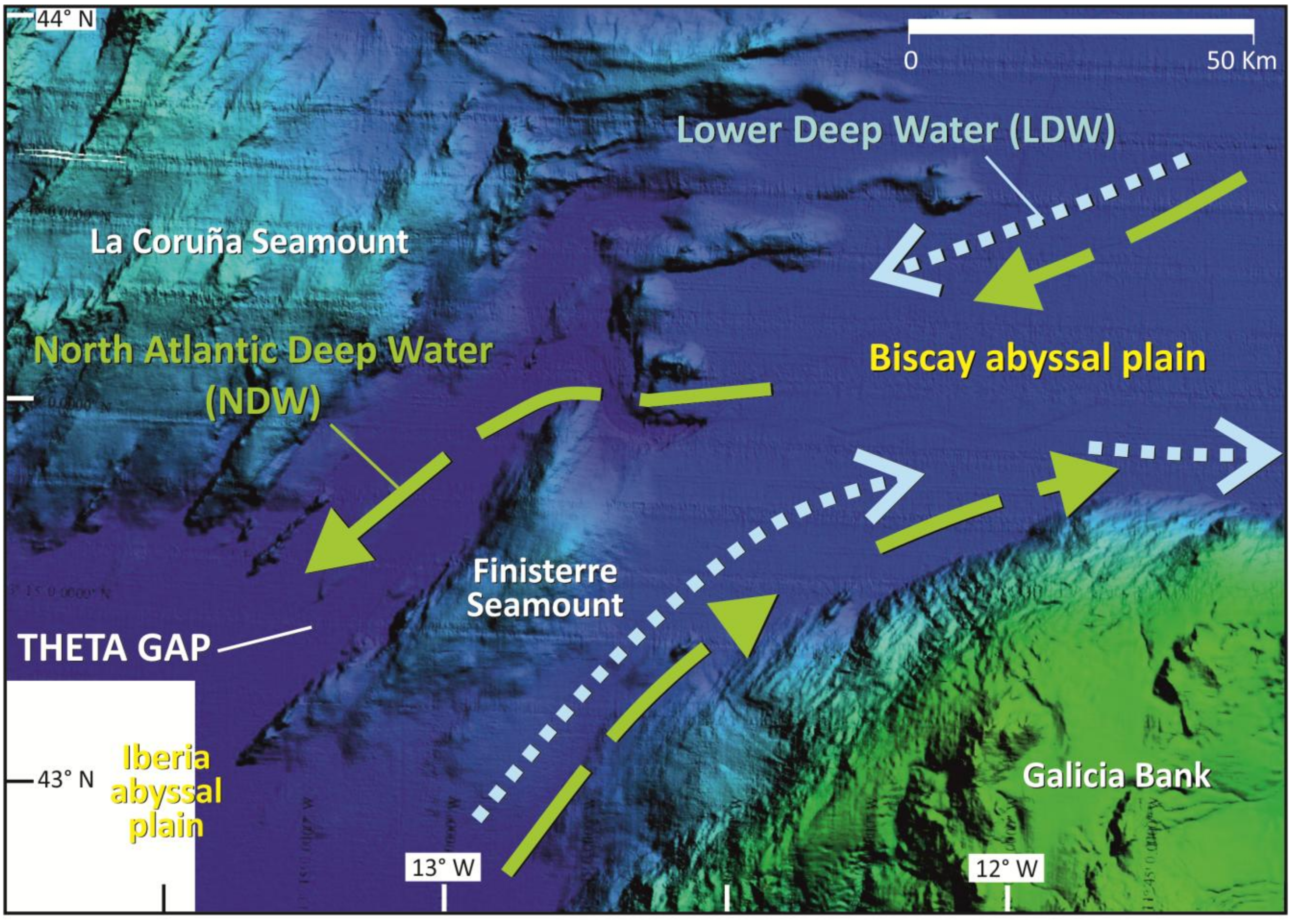

44° N
0
50 Km
Lower Deep Water (LDW)
La Coruña Seamount
North Atlantic Deep Water
(NDW)
Biscay abyssal plain
Finisterre
Seamount
THETA GAP
43° N
Iberia
abyssal
plain
Galicia Bank
13° W
12° W

Fig. 3

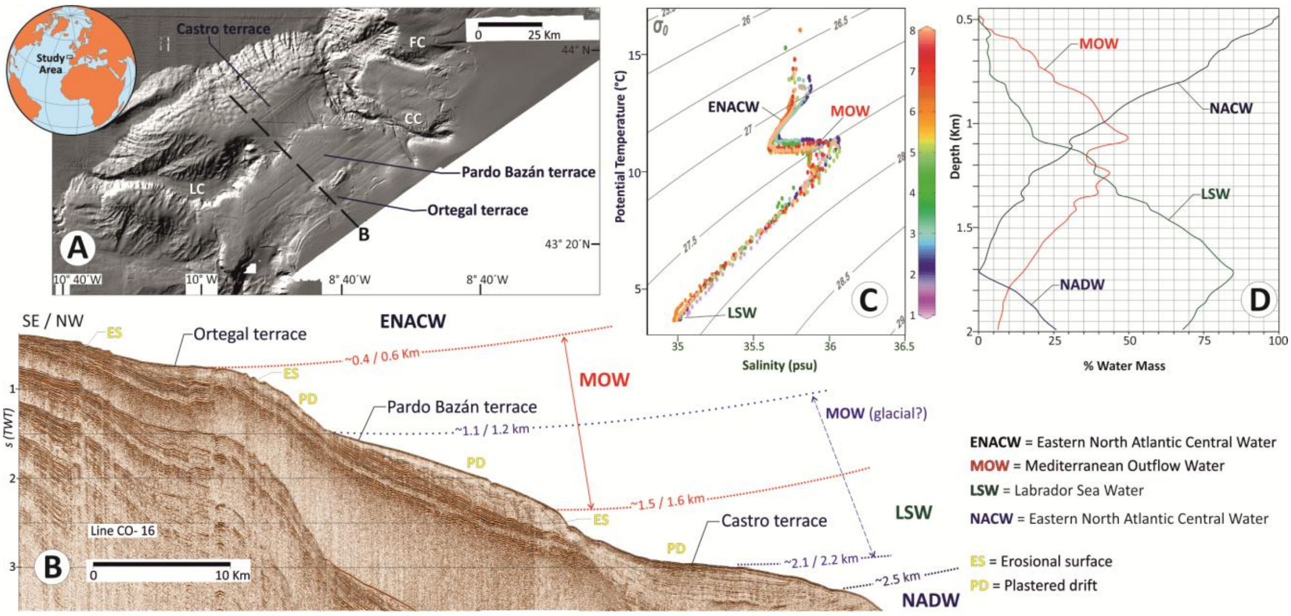

Castro terrace
FC
CC
LC
Pardo Bazán terrace
Ortegal terrace
Study Area
0 25 Km
44° N
43° 20´N
A
B
C
D
ENACW
MOW
LSW
NACW
NADW
SE / NW
Ortegal terrace
Pardo Bazán terrace
Castro terrace
~0.4 / 0.6 Km
~1.1 / 1.2 km
~1.5 / 1.6 km
~2.1 / 2.2 km
~2.5 km
MOW (glacial?)
Line CO- 16
0 10 Km
Salinity (psu)
Potential Temperature (°C)
Depth (Km)
% Water Mass
ENACW = Eastern North Atlantic Central Water
MOW = Mediterranean Outflow Water
LSW = Labrador Sea Water
NACW = Eastern North Atlantic Central Water
ES = Erosional surface
PD = Plastered drift

Fig. 4

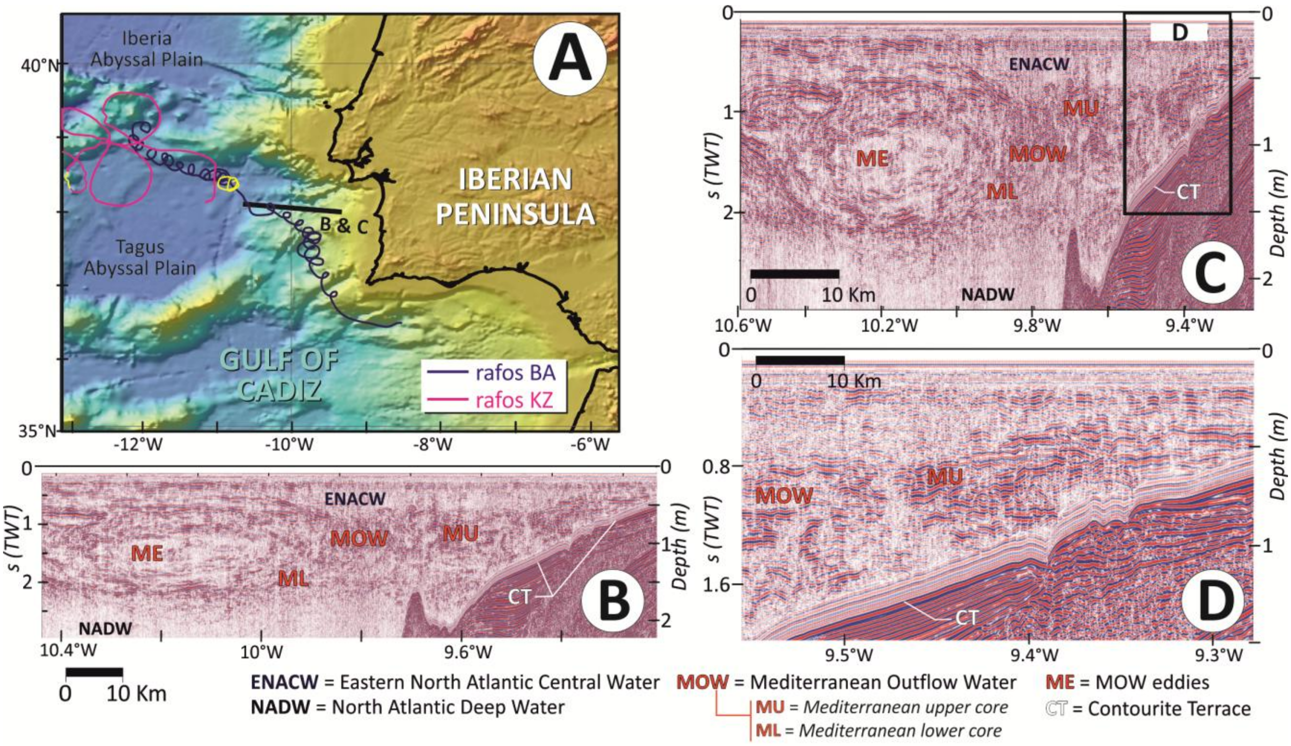

Iberia Abyssal Plain
Tagus Abyssal Plain
IBERIAN PENINSULA
GULF OF CADIZ
B & C
rafos BA
rafos KZ
A
40°N
35°N
-12°W
-10°W
-8°W
-6°W
ENACW
MU
ME
MOW
ML
CT
NADW
C
D
B
s (TWT)
Depth (m)
10 Km
10.6°W
10.2°W
9.8°W
9.4°W
10.4°W
10°W
9.6°W
9.5°W
9.3°W
ENACW = Eastern North Atlantic Central Water
NADW = North Atlantic Deep Water
MOW = Mediterranean Outflow Water
MU = Mediterranean upper core
ML = Mediterranean lower core
ME = MOW eddies
CT = Contourite Terrace

Fig. 5

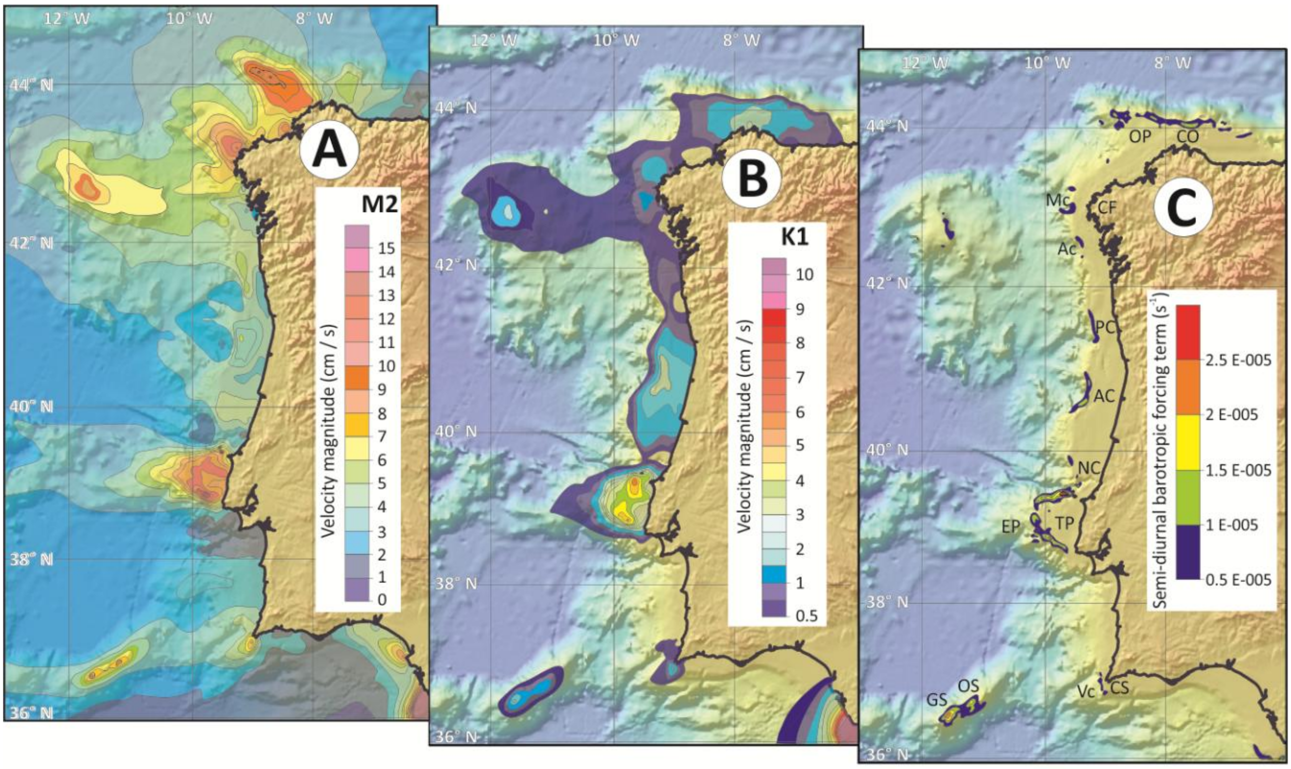

A
M2
Velocity magnitude (cm / s)
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
B
K1
Velocity magnitude (cm / s)
10
9
8
7
6
5
4
3
2
1
0.5
C
Semi-diurnal barotropic forcing term (s-1)
2.5 E-005
2 E-005
1.5 E-005
1 E-005
0.5 E-005
OP
CO
Mc
CF
Ac
PC
AC
NC
EP
TP
GS
OS
Vc
CS
12° W
10° W
8° W
44° N
42° N
40° N
38° N
36° N

Fig. 6

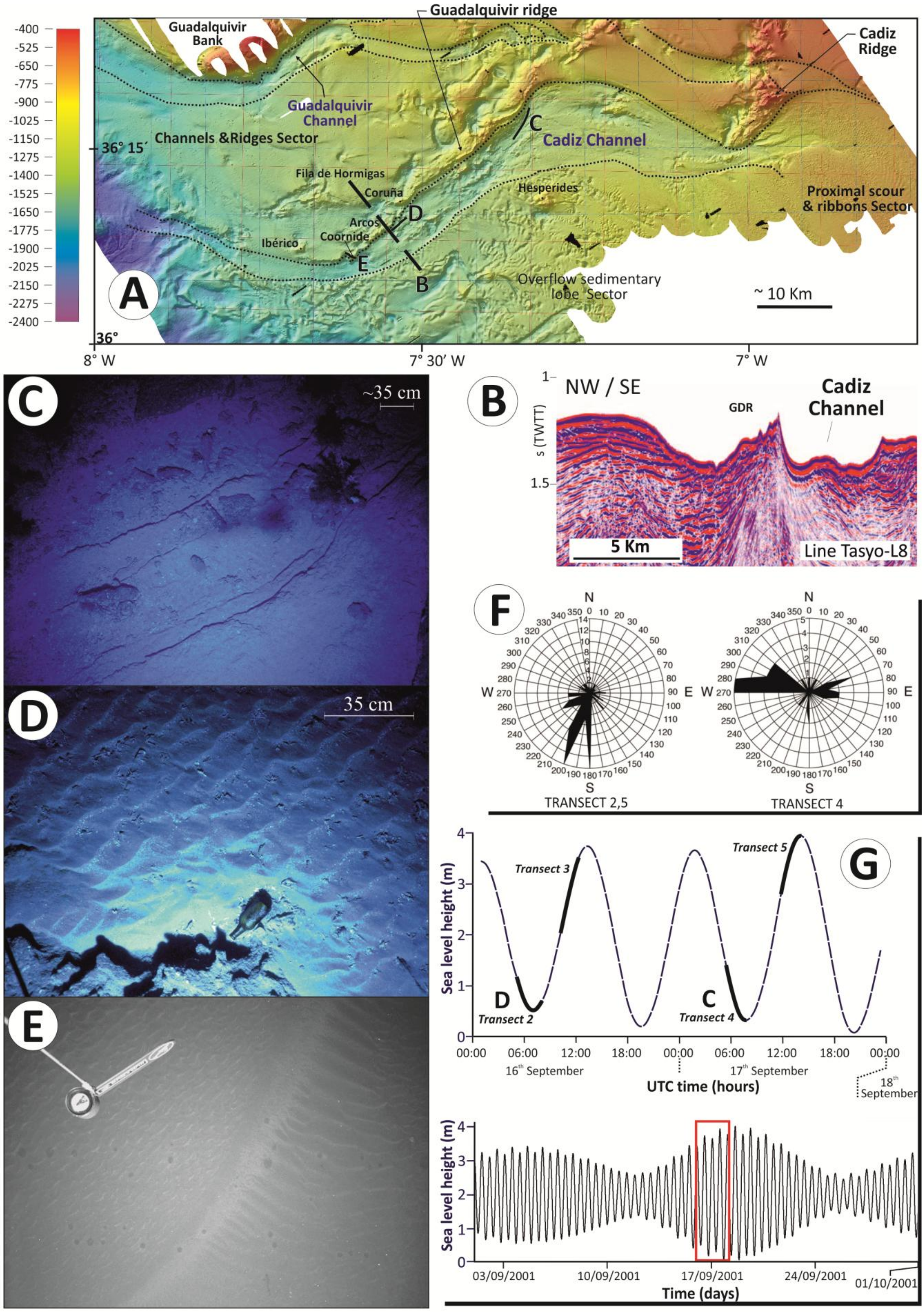

Guadalquivir ridge
Guadalquivir Bank
Cadiz Ridge
Guadalquivir Channel
Channels &Ridges Sector
36° 15´
Cadiz Channel
Fila de Hormigas
Coruña
Hesperides
Proximal scour & ribbons Sector
Arcos
Ibérico
Coornide
Overflow sedimentary lobe Sector
~ 10 Km
36°
8° W
7° 30' W
7° W
-400
-525
-650
-775
-900
-1025
-1150
-1275
-1400
-1525
-1650
-1775
-1900
-2025
-2150
-2275
-2400
~35 cm
35 cm
NW / SE
GDR
Cadiz Channel
s (TWTT)
5 Km
Line Tasyo-L8
TRANSECT 2,5
TRANSECT 4
Transect 3
Transect 5
Transect 2
Transect 4
Sea level height (m)
16th September
17th September
18th September
UTC time (hours)
03/09/2001
10/09/2001
17/09/2001
24/09/2001
01/10/2001
Time (days)

Fig. 7

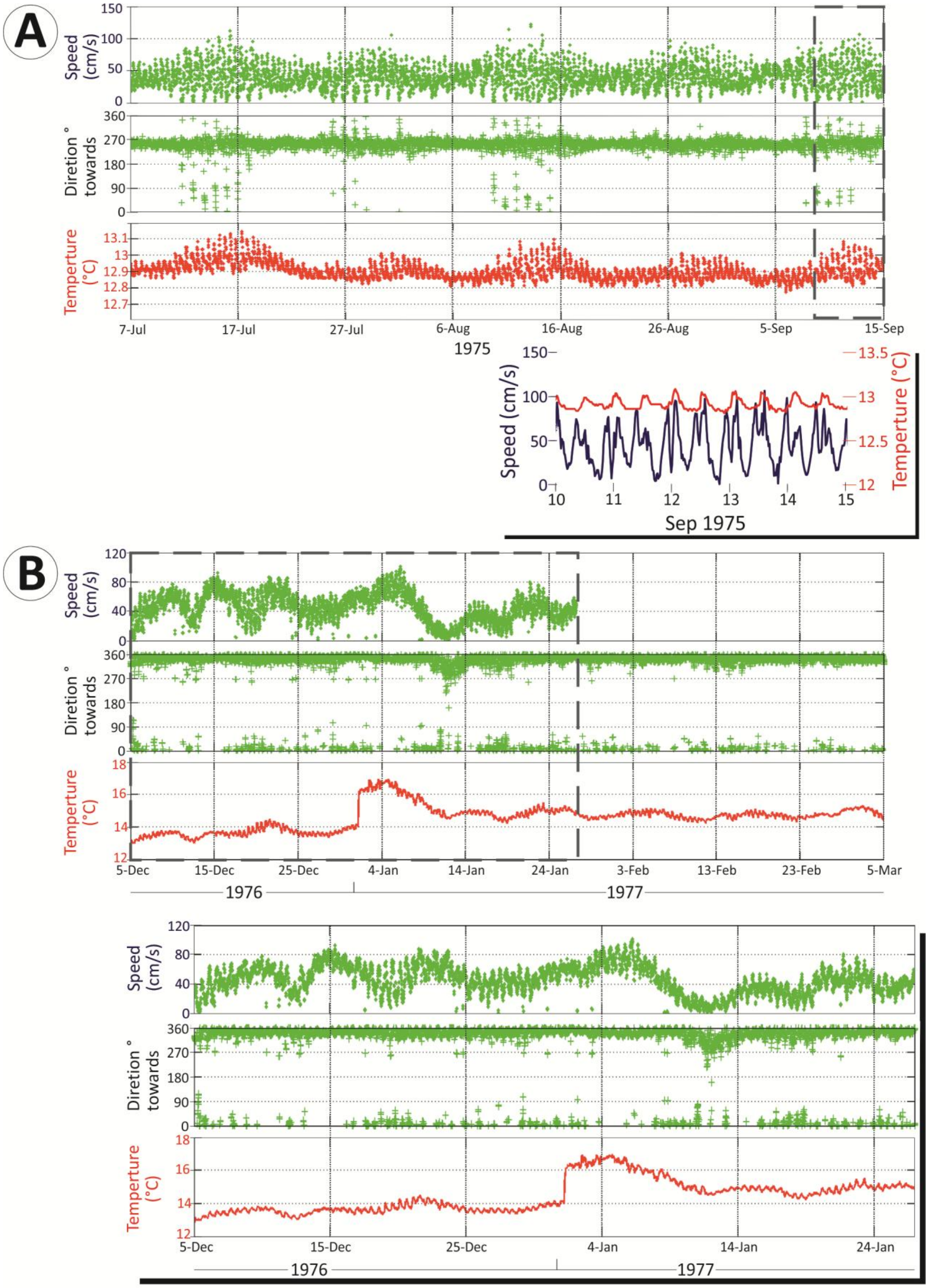

A
Speed (cm/s)
Diretion ° towards
Temperture (°C)
7-Jul
17-Jul
27-Jul
6-Aug
16-Aug
26-Aug
5-Sep
15-Sep
1975
Speed (cm/s)
Temperture (°C)
Sep 1975
B
Speed (cm/s)
Diretion ° towards
Temperture (°C)
5-Dec
15-Dec
25-Dec
4-Jan
14-Jan
24-Jan
3-Feb
13-Feb
23-Feb
5-Mar
1976
1977

Fig. 8

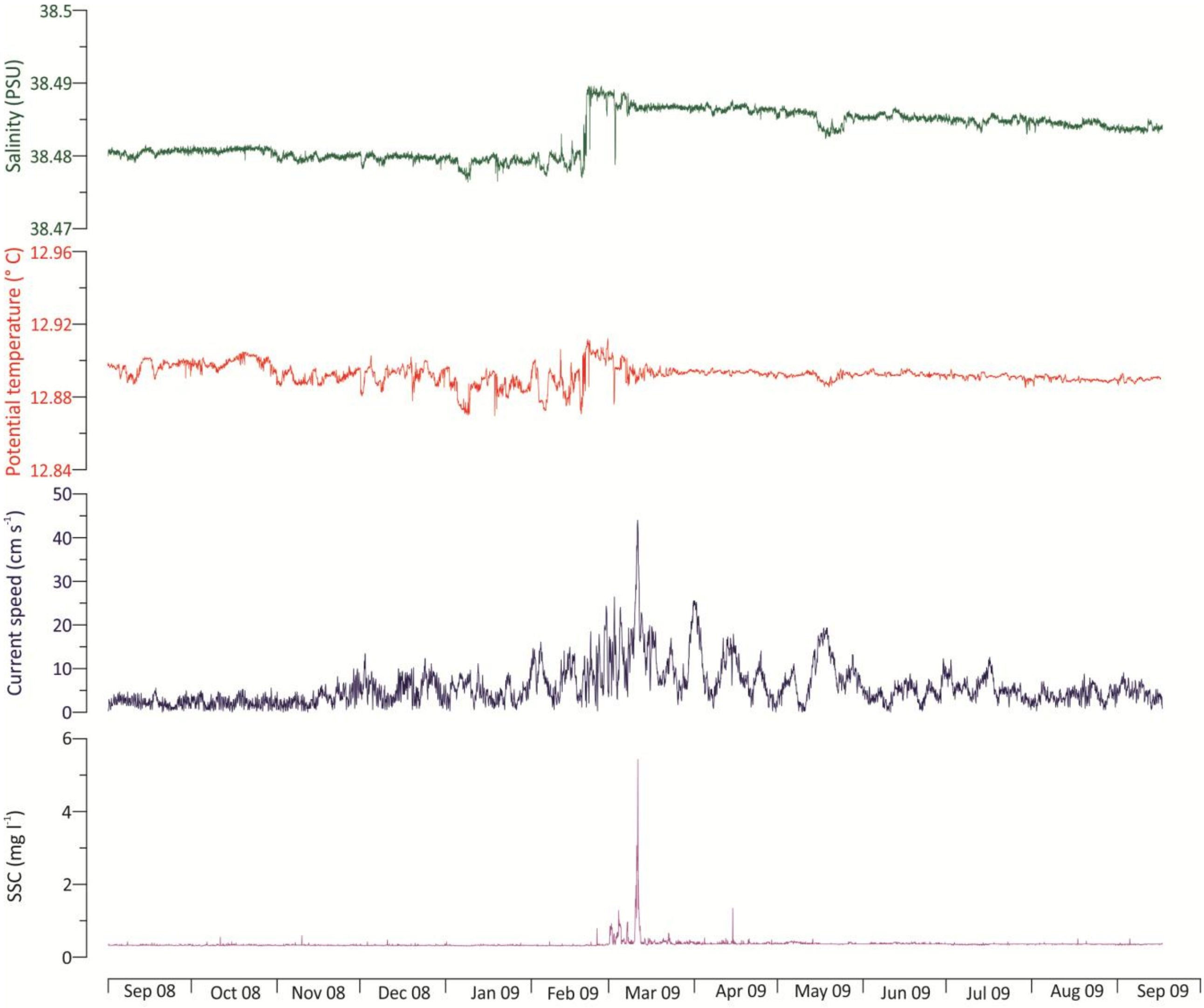

Salinity (PSU)
38.5
38.49
38.48
38.47
Potential temperature (° C)
12.96
12.92
12.88
12.84
Current speed (cm s-1)
50
40
30
20
10
0
SSC (mg l-1)
6
4
2
0
Sep 08
Oct 08
Nov 08
Dec 08
Jan 09
Feb 09
Mar 09
Apr 09
May 09
Jun 09
Jul 09
Aug 09
Sep 09

Fig. 9

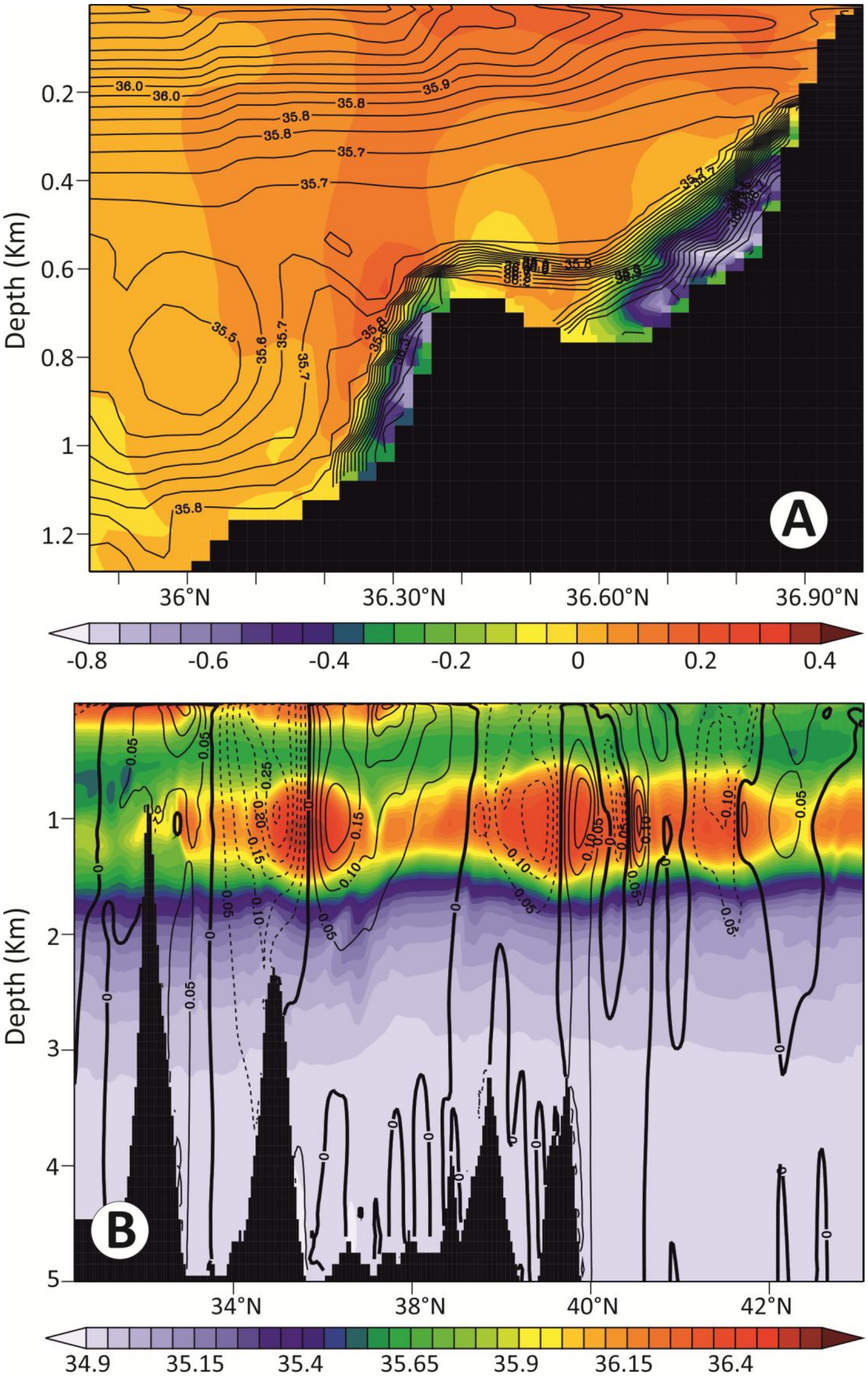

A
Depth (Km)
0.2
0.4
0.6
0.8
1
1.2
36°N
36.30°N
36.60°N
36.90°N
-0.8
-0.6
-0.4
-0.2
0
0.2
0.4
B
Depth (Km)
1
2
3
4
5
34°N
38°N
40°N
42°N
34.9
35.15
35.4
35.65
35.9
36.15
36.4

Fig. 10

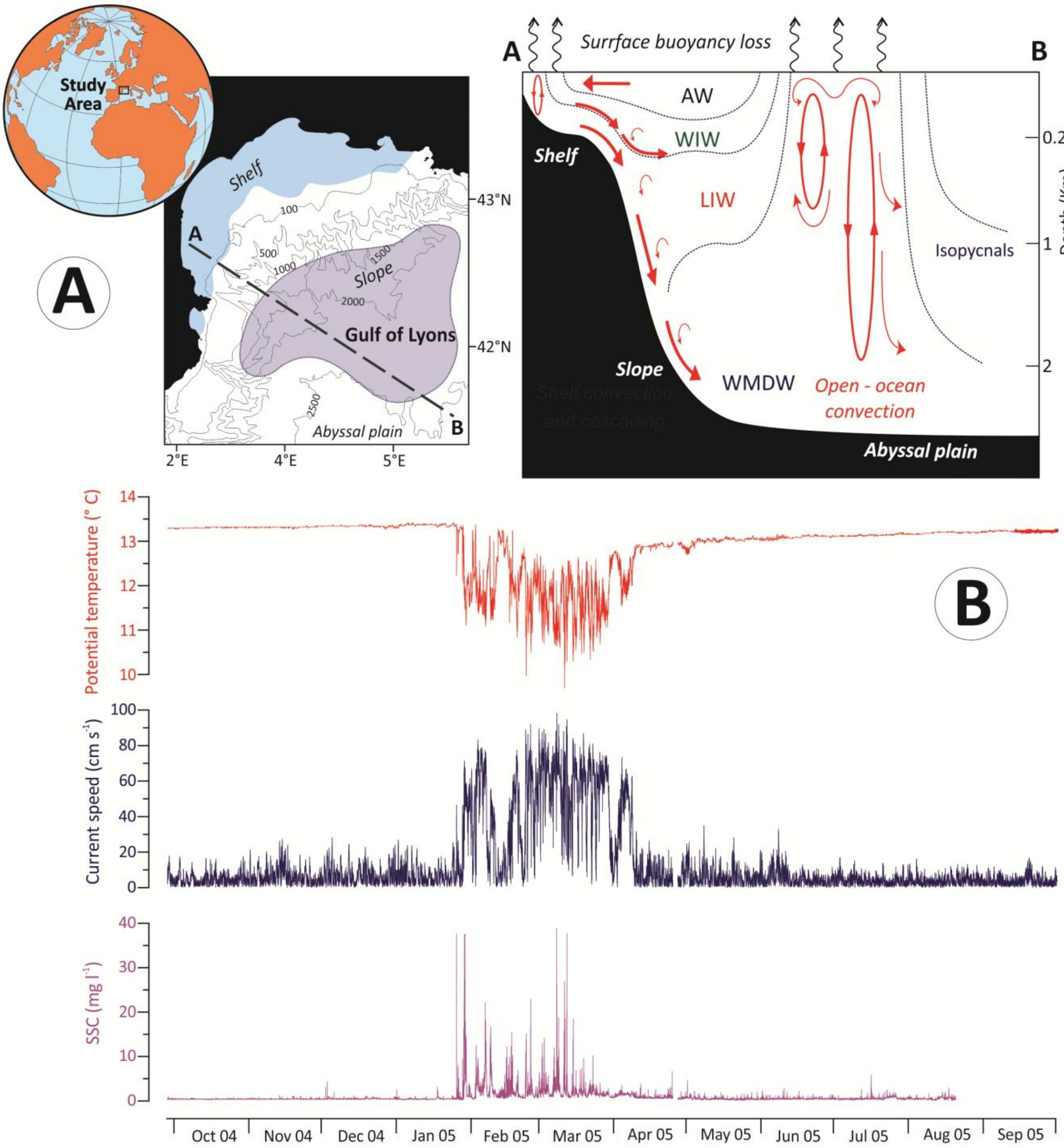

Study Area
Shelf
100
500
1000
1500
2000
Slope
Gulf of Lyons
2500
Abyssal plain
A
B
43°N
42°N
2°E
4°E
5°E
Surrface buoyancy loss
AW
WIW
LIW
WMDW
Shelf
Slope
Shelf convection and cascading
Open - ocean convection
Isopycnals
Abyssal plain
Depth (Km)
0.2
1
2
Potential temperature (° C)
Current speed (cm s-1)
SSC (mg l-1)
Oct 04
Nov 04
Dec 04
Jan 05
Feb 05
Mar 05
Apr 05
May 05
Jun 05
Jul 05
Aug 05
Sep 05

Fig. 11

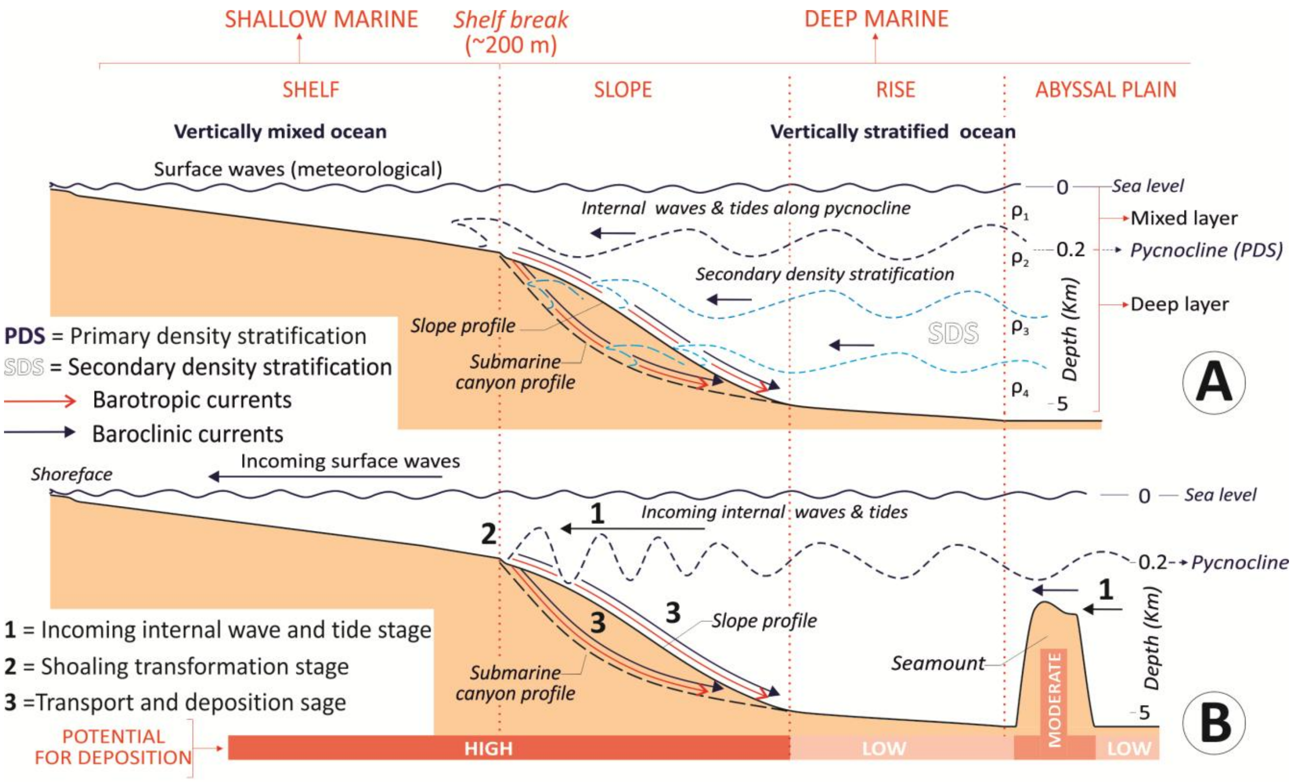

SHALLOW MARINE
Shelf break (~200 m)
DEEP MARINE
SHELF
SLOPE
RISE
ABYSSAL PLAIN
Vertically mixed ocean
Vertically stratified ocean
Surface waves (meteorological)
Internal waves & tides along pycnocline
Secondary density stratification
SDS
Slope profile
Submarine canyon profile
0
Sea level
Mixed layer
0.2
Pycnocline (PDS)
Deep layer
Depth (Km)
5
A
PDS = Primary density stratification
SDS = Secondary density stratification
Barotropic currents
Baroclinic currents
Shoreface
Incoming surface waves
Incoming internal waves & tides
1
2
3
Seamount
B
1 = Incoming internal wave and tide stage
2 = Shoaling transformation stage
3 =Transport and deposition sage
POTENTIAL FOR DEPOSITION
HIGH
LOW
MODERATE

Fig. 12

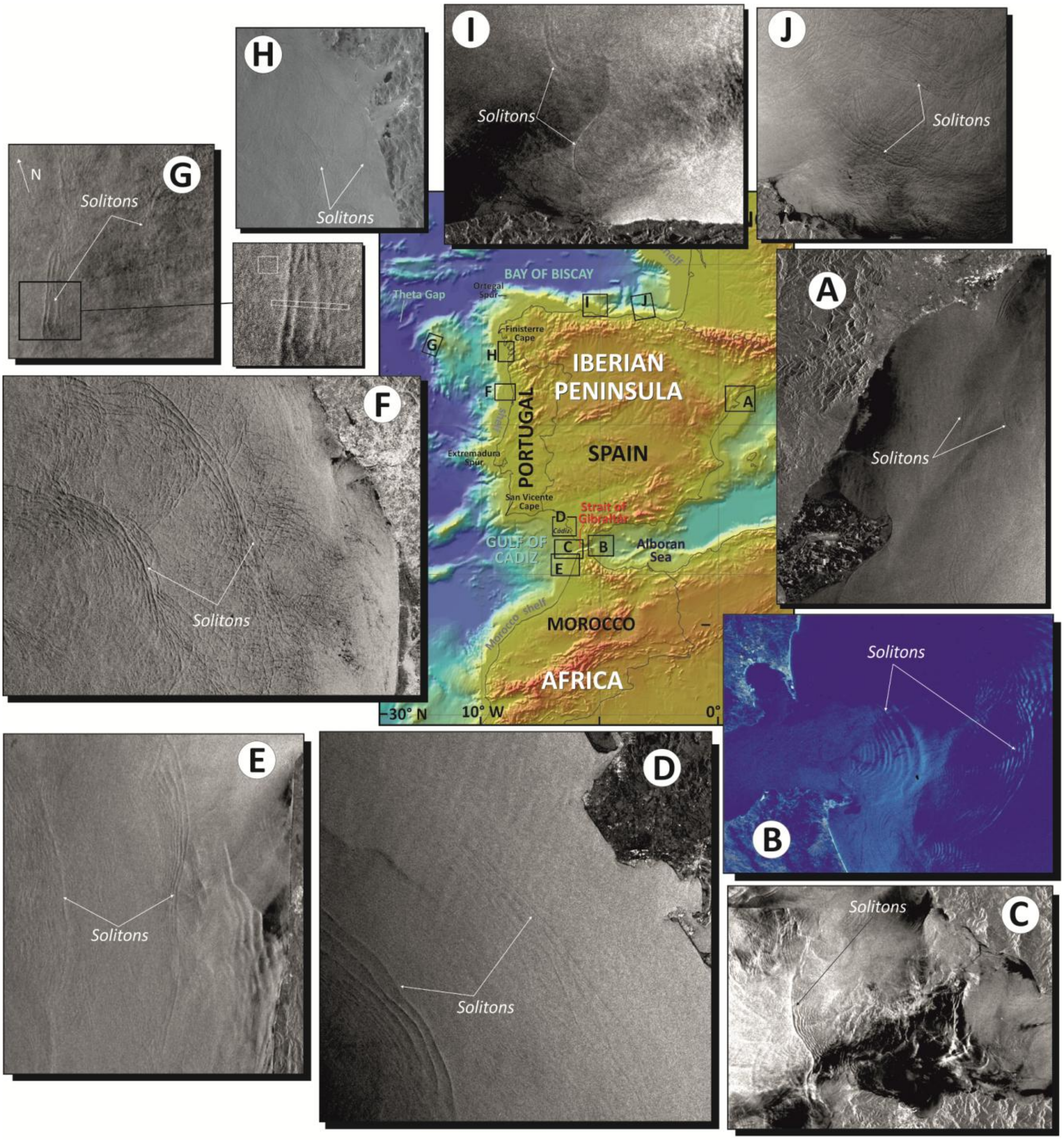

H
Solitons
I
Solitons
J
Solitons
G
N
Solitons
BAY OF BISCAY
Theta Gap
Ortegal Spur
Finisterre Cape
IBERIAN PENINSULA
PORTUGAL
SPAIN
Extremadura Spur
San Vicente Cape
Strait of Gibraltar
Cádiz
GULF OF CADIZ
Alboran Sea
Morocco shelf
MOROCCO
AFRICA
30° N
10° W
0°
A
Solitons
F
Solitons
B
Solitons
E
Solitons
D
Solitons
C
Solitons

Fig. 13

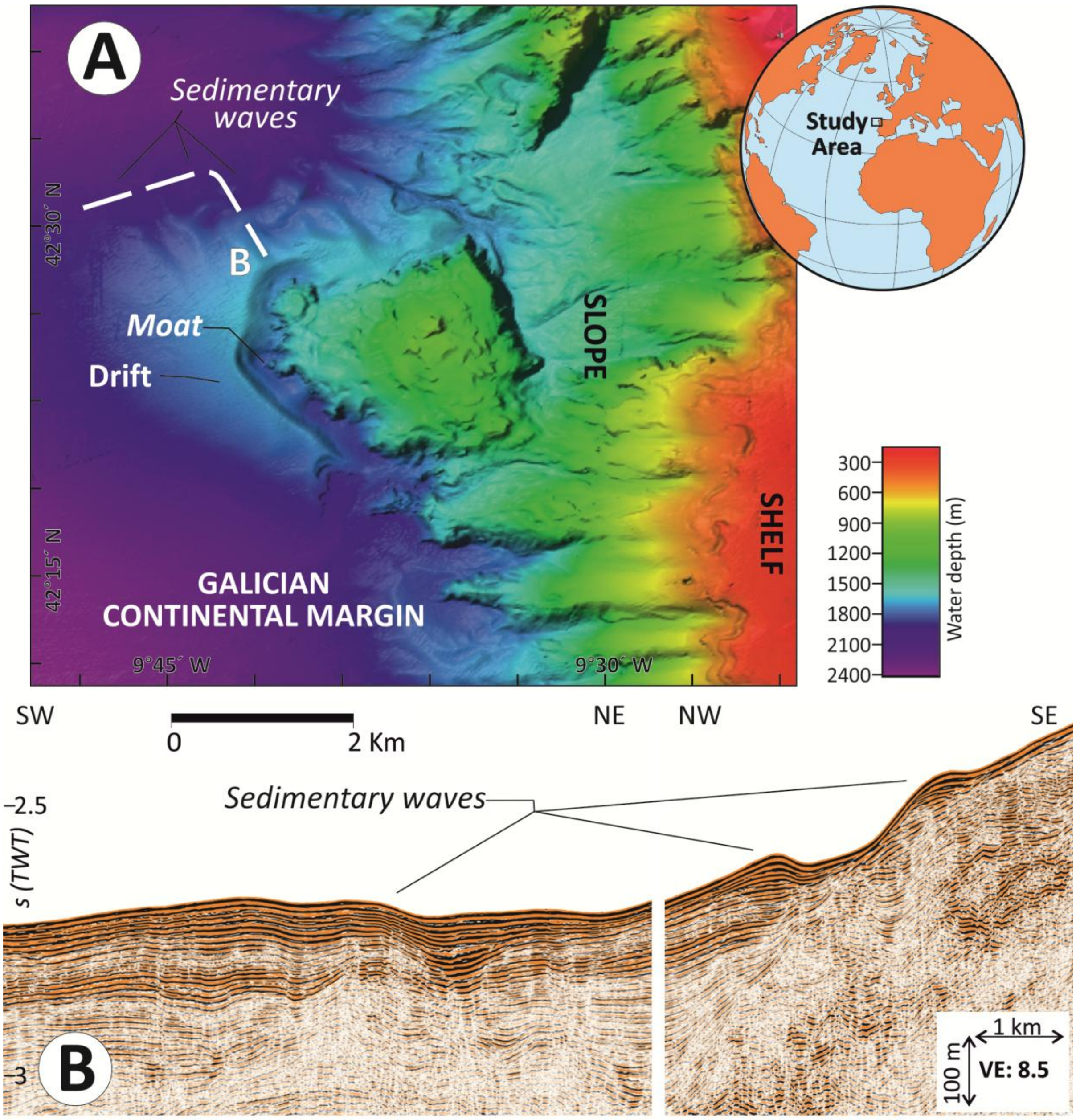

A
Sedimentary waves
B
Moat
Drift
SLOPE
SHELF
GALICIAN CONTINENTAL MARGIN
42°30' N
42°15' N
9°45' W
9°30' W
Study Area
300
600
900
1200
1500
1800
2100
2400
Water depth (m)
SW
NE
NW
SE
0
2 Km
Sedimentary waves
-2.5
-3
s (TWT)
B
1 km
100 m
VE: 8.5

Fig. 14

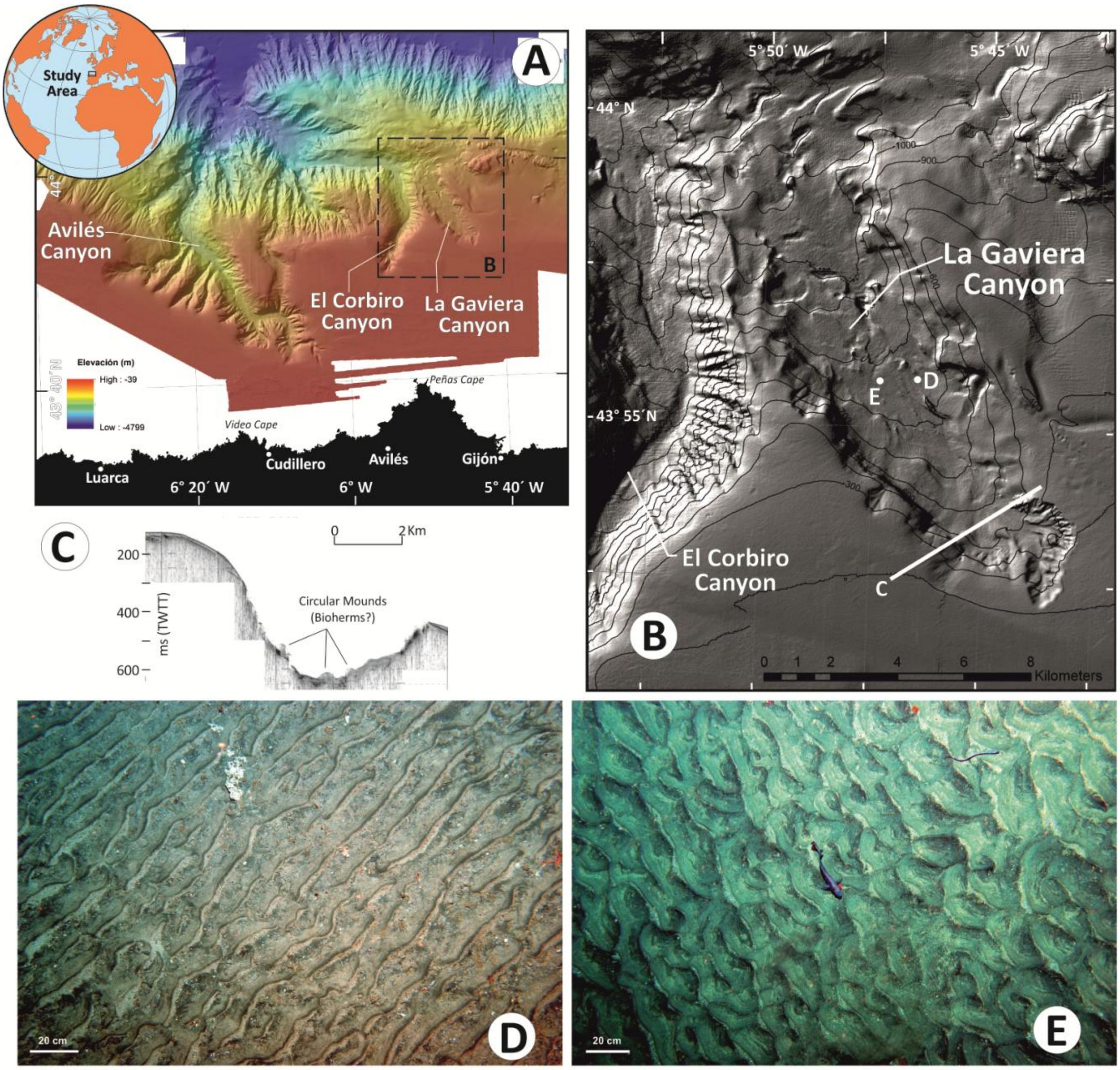

A
Study Area
Avilés Canyon
El Corbiro Canyon
La Gaviera Canyon
B
Elevación (m)
High : -39
Low : -4799
Peñas Cape
Vidio Cape
Luarca
Cudillero
Avilés
Gijón
6° 20´ W
6° W
5° 40´ W
44° N
43° 40´ N
5° 50´ W
5° 45´ W
44° N
43° 55´ N
La Gaviera Canyon
E
D
El Corbiro Canyon
C
B
0 1 2 4 6 8 Kilometers
C
200
400
600
ms (TWTT)
0 2 Km
Circular Mounds (Bioherms?)
20 cm
D
20 cm
E

Fig. 15

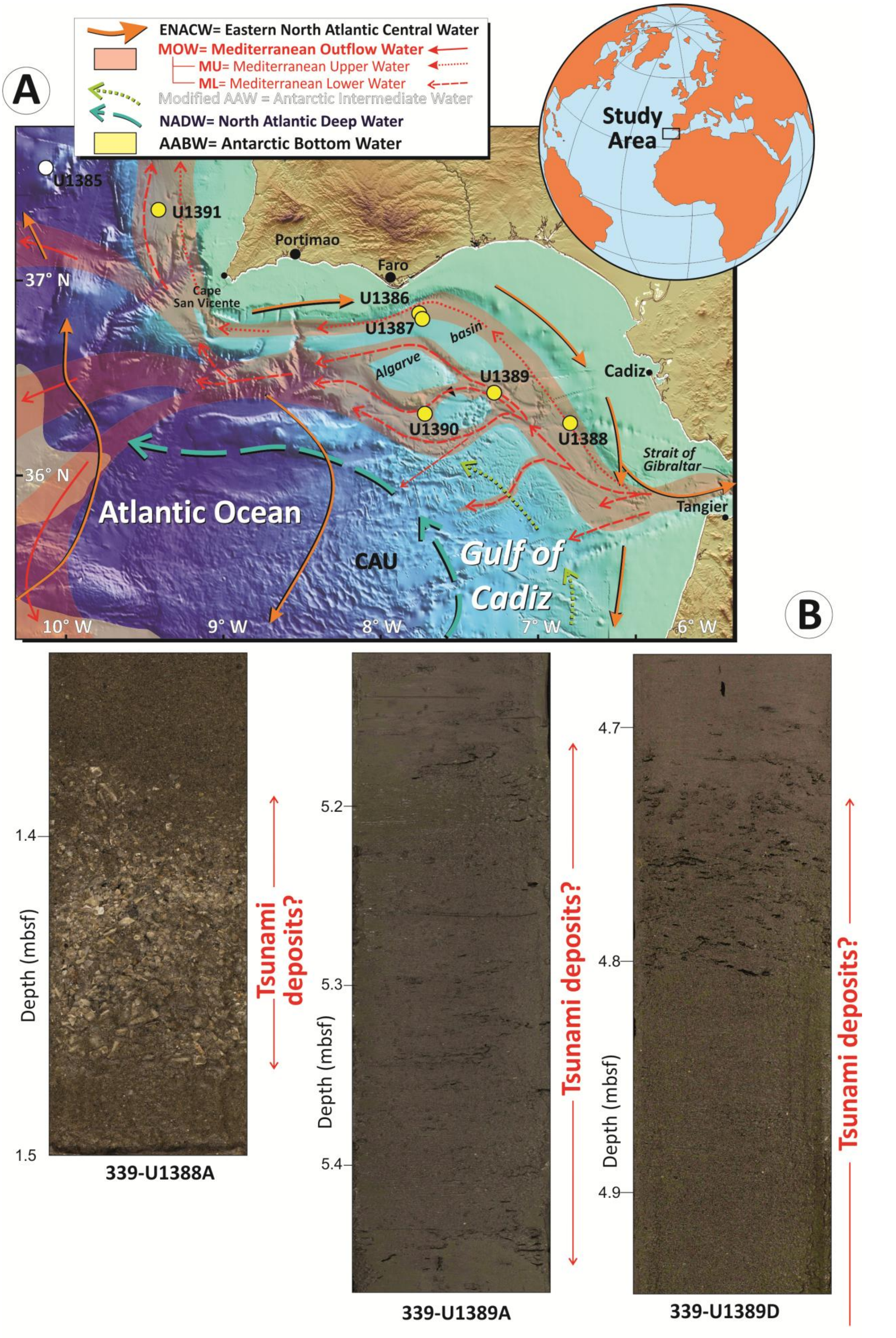

A
ENACW= Eastern North Atlantic Central Water
MOW= Mediterranean Outflow Water
MU= Mediterranean Upper Water
ML= Mediterranean Lower Water
Modified AAW = Antarctic Intermediate Water
NADW= North Atlantic Deep Water
AABW= Antarctic Bottom Water
Study Area
U1385
U1391
Portimao
Faro
Cape San Vicente
U1386
U1387
basin
Algarve
U1389
Cadiz
U1390
U1388
Strait of Gibraltar
Tangier
37° N
36° N
Atlantic Ocean
CAU
Gulf of Cadiz
10° W
9° W
8° W
7° W
6° W
B
1.4
1.5
Depth (mbsf)
Tsunami deposits?
339-U1388A
5.2
5.3
5.4
339-U1389A
4.7
4.8
4.9
339-U1389D

Fig. 16

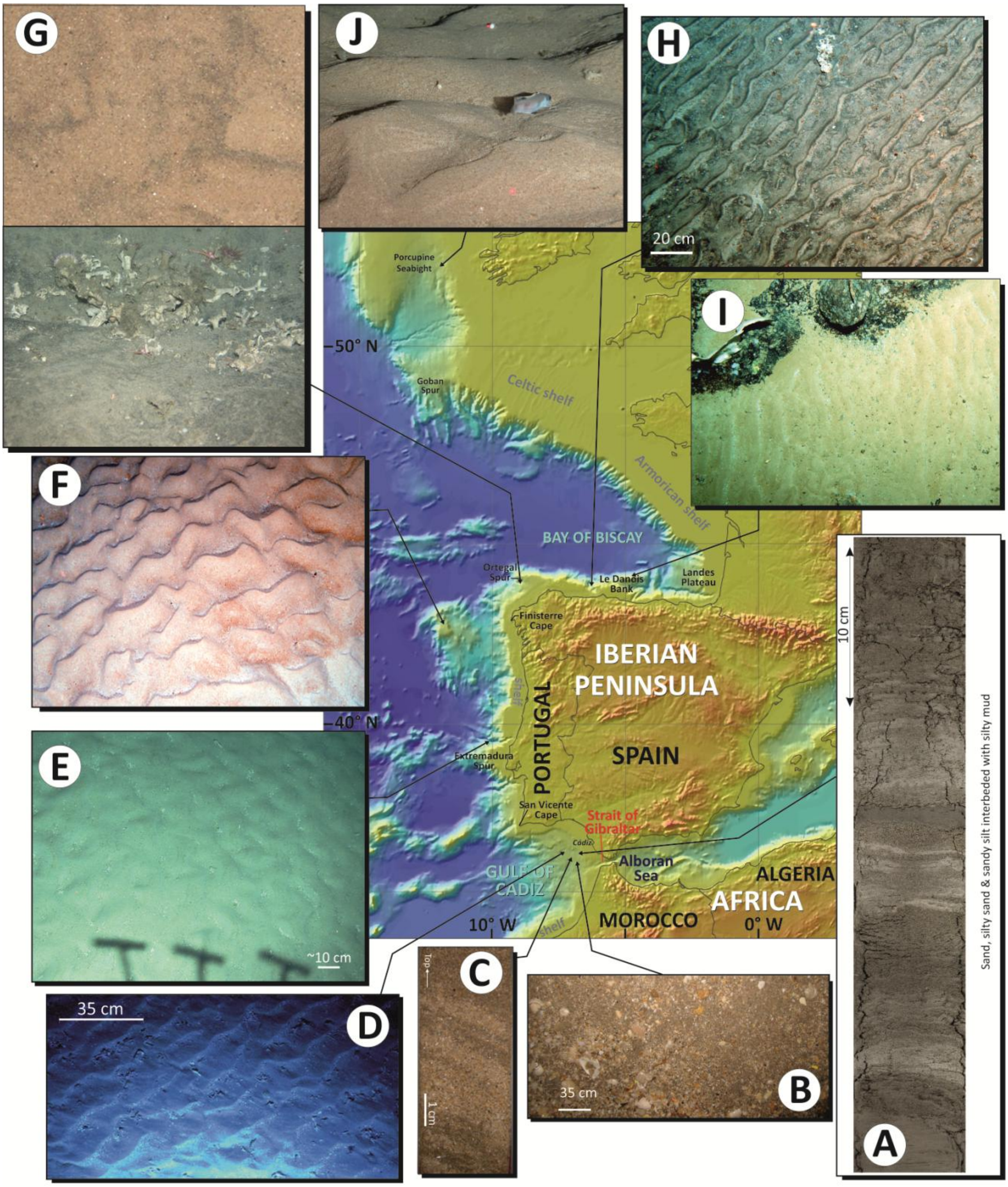

G
J
H
20 cm
I
F
E
~10 cm
D
35 cm
C
Top
1 cm
B
35 cm
A
10 cm
Sand, silty sand & sandy silt interbeded with silty mud
Porcupine Seabight
50° N
Goban Spur
Celtic shelf
Armorican shelf
BAY OF BISCAY
Ortegal Spur
Le Danois Bank
Landes Plateau
Finisterre Cape
IBERIAN PENINSULA
shelf
PORTUGAL
40° N
SPAIN
Extremadura Spur
San Vicente Cape
Strait of Gibraltar
Cádiz
Alboran Sea
ALGERIA
GULF OF CADIZ
AFRICA
10° W
MOROCCO
0° W

Fig. 17

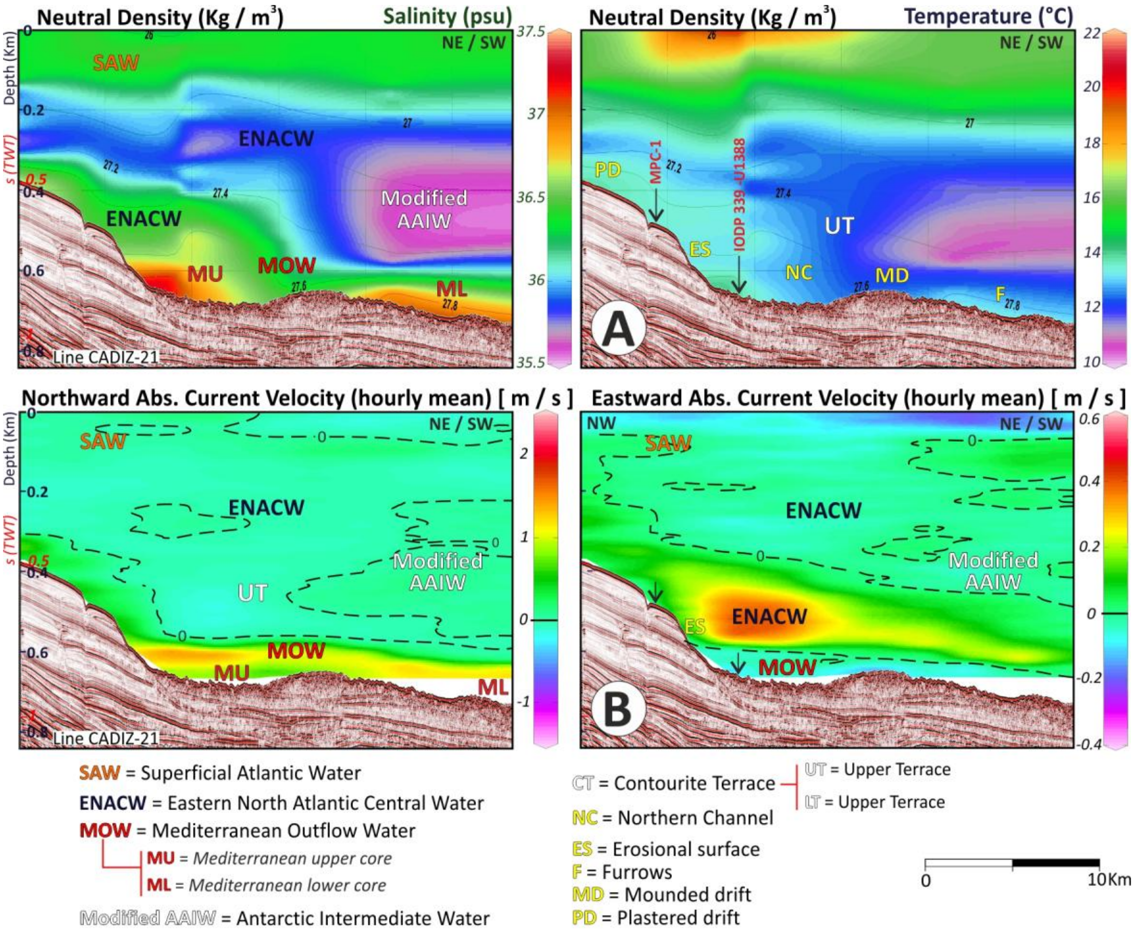

Neutral Density (Kg / m3)
Salinity (psu)
Neutral Density (Kg / m3)
Temperature (°C)
Northward Abs. Current Velocity (hourly mean) [m / s]
Eastward Abs. Current Velocity (hourly mean) [m / s]
NE / SW
Depth (Km)
s (TWT)
Line CADIZ-21
MPC-1
IODP 339 -U1388
A
B
SAW = Superficial Atlantic Water
ENACW = Eastern North Atlantic Central Water
MOW = Mediterranean Outflow Water
MU = Mediterranean upper core
ML = Mediterranean lower core
Modified AAIW = Antarctic Intermediate Water
CT = Contourite Terrace
UT = Upper Terrace
LT = Upper Terrace
NC = Northern Channel
ES = Erosional surface
F = Furrows
MD = Mounded drift
PD = Plastered drift
0
10Km

Fig. 18

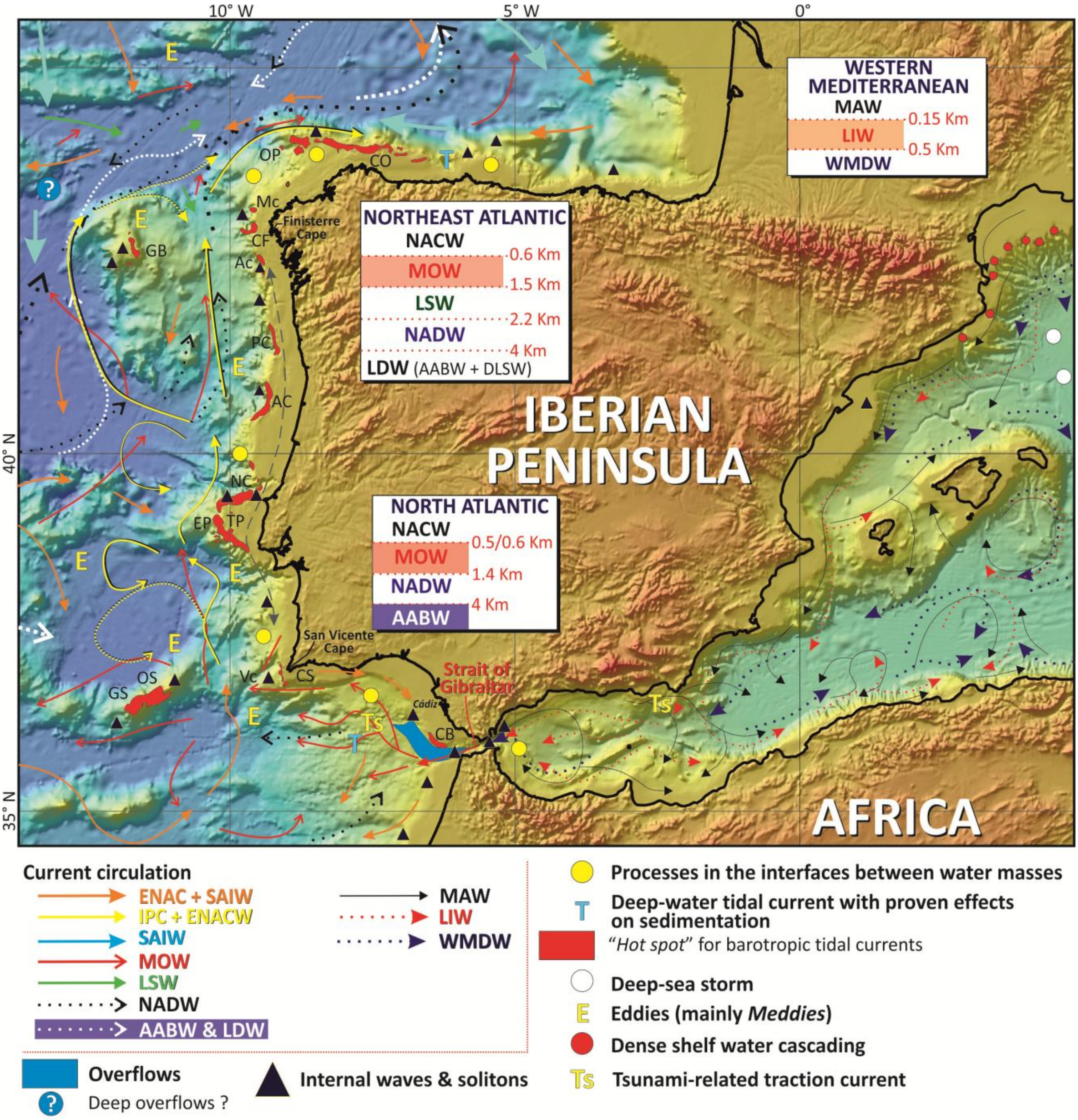
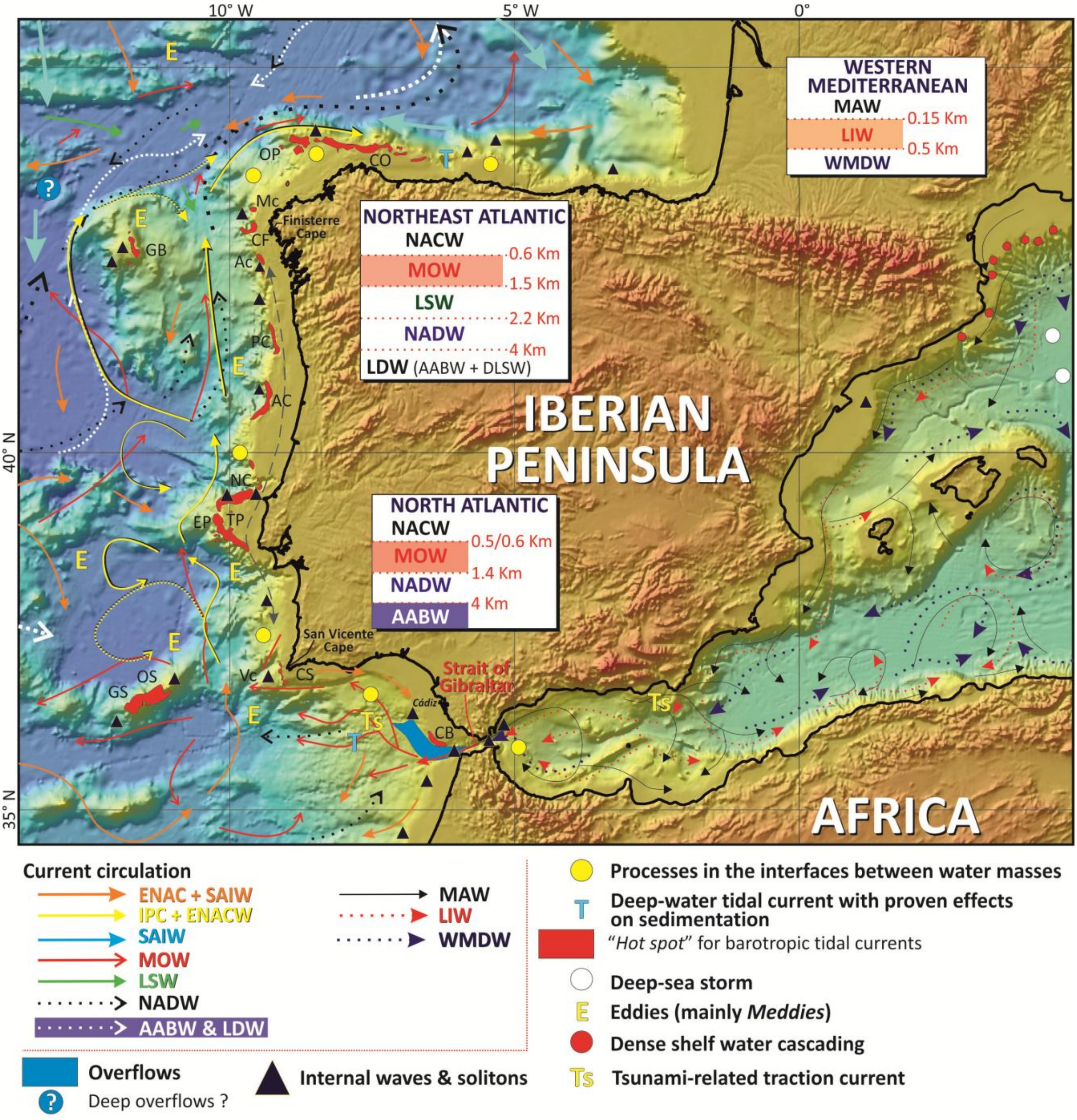

10° W
5° W
0°
40° N
35° N
WESTERN MEDITERRANEAN
MAW
0.15 Km
LIW
0.5 Km
WMDW
NORTHEAST ATLANTIC
NACW
0.6 Km
MOW
1.5 Km
LSW
2.2 Km
NADW
4 Km
LDW (AABW + DLSW)
NORTH ATLANTIC
NACW
0.5/0.6 Km
MOW
1.4 Km
NADW
4 Km
AABW
IBERIAN PENINSULA
AFRICA
Finisterre Cape
San Vicente Cape
Strait of Gibraltar
Cádiz
OP
CO
Mc
CF
Ac
GB
PC
AC
NC
EP
TP
CS
Vc
OS
GS
CB
E
T
Ts
Current circulation
ENAC + SAIW
IPC + ENACW
SAIW
MOW
LSW
NADW
AABW & LDW
MAW
LIW
WMDW
Processes in the interfaces between water masses
Deep-water tidal current with proven effects on sedimentation
"Hot spot" for barotropic tidal currents
Deep-sea storm
Eddies (mainly Meddies)
Dense shelf water cascading
Tsunami-related traction current
Overflows
Deep overflows ?
Internal waves & solitons

Fig. 19

Oceanographic processes	Main sedimentary processes	Main morphosedimentary products	Examples along the Iberian margins
Bottom currents & thermohaline circulation	Erosion; Sed. Transport/Reworked; Deposition	Large contourite deposits (*drifts*) & erosional features	*NE Iberian margin; Alboran Sea; Gulf of Cadiz; WIM; Galician margin; Cantabrian Sea*
Overflows	Erosion; Sed. Transport; Depositional (locally and distally)	Contourite terraces; large channels, scours, furrows; sandy-sheeted deposits, ripples, dunes & sand/gravel ribbons	*Strait of Gibraltar; Theta Gap (?)*
Processes at the interface between water masses (waves & eddies)	Erosion; Sed. Transport (re-suspension); Deposition	Morphological changes along the slope gradient; large contourite terraces, erosional abraded surfaces	*Alboran Sea, Gulf of Cádiz; Portuguese Margin; Galician margin and Cantabrian Sea*
Deep-water tidal currents	On submarine canyons and adjacent areas, contourite channels: Reworked and deposition	Bedforms (ripples & dunes)	*Atlantic margins (especially at the Extremadura and Ortegal Spurs and Gulf of Cadiz)*
Deep-sea storms	Erosion; Sed. Transport (re-suspension)	Erosional & depositional	*Ubiquitous (?).Atlantic & Mediterranean margins*
Eddies	Erosion; Sed. Transport (re-suspension)	Erosional & depositional	*Gulf of Cadiz; West off Portugal; Galician margin*
Secondary circulation	Erosion	Scours; furrows	*Ubiquitous. Atlantic & Mediterranean margins*
Dense shelf water cascades	Sed. Transport (density-driven flow, re-suspension)	Erosional & depositional	*Mediterranean margins (Gulf of Lions)*
Internal waves & solitons	Erosional; Sed. Transport/Reworked; Deposition	Erosional & depositional Dunes; Sedimentary waves	*Ubiquitous. Internal waves along the margins. Solitons associated to shelf breaks*
Other-related traction currents (tsunamis, rogue- & cyclone waves)	Sed. Transport (mud flow, re-suspension); Slope instabilities	> Sed. Rates; Coarser sed. input	*Southern Iberian margin, Alboran Sea*

Table-I.- Summary of the oceanographic processes and products along the Iberian margins.

Graphical abstract



Highlights

- Knowledge of bottom-currents and associated processes is rapidly evolving.
- Deep-water processes generate depositional and erosional features.
- Interfaces between water masses control the contourite terraces development on slopes.
- Contourites deposits exhibit greater variation than the established facies model.
- A new multidisciplinary approach in future contourite research is needed.